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Radiation fog: investigations at Cardington, 1951-54

by

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Radiation fog: investigations at Cardington, 1951 - 54

1. Introduction

Measurements of temperature, humidity and wind speed in the lowest 4000 ft. of the atmosphere have been made at Cardington since 1944, using the balloon-borne thermocouple psychrometer and cup anemometer described by Swinbank (1946). Routine measurements have been made with fair regularity three times a day (0700, 1200, 1800 hr.) except in strong winds. In the autumn of 1951 these balloon soundings were made the basis of a considerably expanded programme of observations at Cardington, in the hope of learning more about the processes of fog formation. This report reviews the results of the expanded programme and the contribution they make to fog problems; it does not attempt to deal with the great body of data available from the ten years of routine observations.

1.1. Observations and apparatus

The general plan of the investigation was to make regular measurements of as many as possible of the quantities that might be important in fog formation on as many as possible of the occasions when radiation fog was to be expected. In addition to the balloon soundings mentioned above and the usual hourly observations of a synoptic reporting station, it was planned to make the following measurements:-

(a) Continuous recordings of "visibility" (atmospheric attenuation coefficient for visible light) by photoelectric meters at 4ft. and 55ft. above ground level.

(b) Height of top of fog, by a photoelectric instrument carried on the balloon.

(c) Earth temperatures at several depths down to 40 cm.

(d) "Surface temperature" as indicated by a spirit thermometer.

(e) Dewfall, by weighing methods.

(f) Atmospheric pollution, by filter-paper gauge.

(g) Wind speed, by sensitive cup anemometers } at height of 4ft., 20ft. and 55ft. on a steel tower

(h) Temperature } by aspirated thermocouple psychrometers } (continuous recordings)

(j) Humidity } by aspirated thermocouple psychrometers } (continuous recordings)

(k) Size and number of water drops in fog, by impactor technique.

(l) Specific heat of soil samples.

(m) "Slant visibility" from the ground to various heights.

Details of the apparatus used in these measurements are given in Appendix I. It has, unfortunately, proved impossible to carry out all the measurements planned; many of the instruments have been difficult to instal and maintain, and the limited staff available (1 E.O. or S.E.O. with 2 or 3 assistants) have not had time to keep them all in working order. First priority has been given to the routine daily balloon soundings and to making frequent soundings on occasions when fog was forecast. In addition, measurements (a) - (g) have been made with fair regularity, as will be seen from the results given below, but (h) and (j) have not yet given reliable results and (k) and (l) have not been attempted at all. (m) was not added to the programme till 1953; measurements have since been made on several nights, but not on all nights, because they interrupt the other observations.

1.2. Occasions on which observations were made

Detailed observations (i.e. hourly synoptic observations, balloon

/soundings

soundings at 2 - 3 hr. intervals and as many as possible of the measurements (a) - (m) mentioned above) were begun at midday on any day when the Central Forecast Office forecast the possibility of fog or clear skies at night and were continued until the fog cleared, until the temperature began to rise after dawn, if no fog had formed, or until it became clear that the forecast was incorrect. Observations were made in the winter months only (Sept. - Mar. inclusive). They have been made on a total of 60 nights in the past three seasons. About 30 suitable nights have been missed, chiefly because of unserviceability of the balloon or lack of a balloon crew. Many of the nights worked had, in fact, considerable amounts of cloud; for simplicity, they are not considered in this report (they include 9 nights with low stratus cloud, which might well repay separate investigation). The remaining nights are made up as follows: 18 with thick fog; 5 with thin fog (visibility < 1100 yd. but > 220 yd.); 11 clear nights (average cloud less than 2/8 medium and low or 4/8 high).

1.3. Discussion of results; plan of the report

The results of the investigation are discussed below under three headings. First there is the general picture of events on a foggy night given by the Cardington observations; this is not, in essentials, different from pictures based on previous work, such as the Leafield and Rye observations, but it is in some respects more detailed, and an attempt has been made to interpret the picture as fully as possible in terms of physical processes. The second heading is the detailed quantitative discussion of these physical processes in the light of the data available. Finally the new data can be used in attempts to find or test semi-empirical rules or formulae for fog forecasting.

The new Cardington data appear to be reasonably satisfactory for work under the first heading - in fact it is suggested that further collection of general data of the same type is likely to yield rapidly diminishing returns. For work under the second and third headings, however, the new data seem inadequate; for the study of physical processes the need is for more detailed, accurate and carefully planned measurements, while for the proper testing of forecasting procedures it is essential to have data for many more occasions. The report therefore ends with recommendations for changes in the plan of the research intended to produce more useful results.

It is clearly almost impossible to be certain that all possible information has been extracted from a large quantity of measurements such as is available from Cardington. It is hoped that the present report has covered the most profitable lines of inquiry and, perhaps more important, that it will give a good idea of the scope and limitations of the results obtained and thus serve as a basis for discussion of possible further analysis or fuller presentation of the data.

2. General qualitative description and discussion of results on clear and foggy nights

The behaviour of temperature and humidity varies so much from night to night that much of the value of the observations is lost by attempting to summarize them in the form of statistics. It seems better to present and discuss observations made on a few selected nights, with references to mean values observed on other occasions as appropriate.

Figs. 1 - 6 show the results obtained on 6 nights selected, as far as possible, to illustrate the main types of behaviour observed. The diagrams show, in their lower parts, the profiles of temperature, dew point and wind speed obtained from the balloon soundings; the temperature profiles have been extended below ground level by using earth thermometer readings. Above these profiles is shown the time-variation of elements observed at screen and ground level. In most cases the graphs are drawn through points representing average values observed over a few minutes at each hour. The

original records - from visibility meters or anemometers for example- often give much greater detail.

2.1. Brief description of typical cases

The main features of figs. 1 - 6 and the occasions they represent will be described briefly, as a preliminary to a discussion of some general characteristics of the results; a more detailed description and discussion of the individual cases follows in section 2.3.

Figs. 1 and 2 represent the commonest and simplest type of behaviour resulting in fog formation. The temperature profiles in the early evening show a small but steep inversion below 50 ft. with a lapse tending towards the dry adiabatic value at greater heights. During the night the air cools rapidly near the ground but more slowly higher up, so that the inversion near the ground becomes steeper and extends to greater heights. When the temperature near the ground has fallen far enough, fog forms in a somewhat thin, shallow and patchy layer; at this stage the fog appears to have little effect on the temperature structure, but after some time the fog becomes denser and deeper, the ground surface temperature rises sharply and the temperature gradient within the fog layer becomes zero or negative. Above the fog there is a steep inversion, similar to the one near the ground before fog formed, changing gradually to a lapse at greater heights.

Rather more than half the foggy nights studied follow the pattern of events shown in figs. 1 and 2 fairly closely. The remainder show a wide variety from which it is difficult to pick out recognizable types of behaviour of frequent occurrence, but figs. 3 and 4 have been selected to show patterns that are followed, at least approximately, on a number of occasions, particularly during prolonged spells of fog. Their common feature is that cooling in the lowest few hundred feet is more or less uniform instead of being greatest near the ground, as in figs. 1 and 2. In fig 3 this is clearly connected with the fact that fog formed early in the night - in fact the chief interest in this example is in showing how a fog grows and changes during the night. In fig. 4 the obvious point of difference from earlier examples is the marked inversion, with dry air above, at about 1500 ft. which is to be attributed to subsidence. Such an inversion is almost always accompanied by relatively uniform cooling in the lower layers and by rapid increase in depth of the fog, once it has formed, as shown in fig. 4. Figs. 5 and 6 show two occasions typical of those on which fog did not form. In fig. 5 there is considerable cooling similar to that shown in figs. 1 and 2, but, presumably because of the rather low initial relative humidity, no fog formed. Fig. 6 is typical of many occasions in winter with temperatures in the region of freezing point, when the total amount of cooling is small and no fog, or at most only very thin fog, forms.

2.2. Advective effects

In the above brief summary the diagrams have been classified chiefly by reference to the temperature profiles. Although there are many minor irregularities, these profiles do follow a fairly regular pattern, with maximum cooling near the ground and less marked cooling aloft. The dew point profiles are very different; it is true that there is a general tendency for dew point to decrease with time below about 500 ft. but above that height no regular pattern is discernible. This general impression from figs. 1 - 6 is confirmed by the summary of changes observed on all foggy or clear nights in fig. 7. Temperature changes (from 1800 - 0600 hr.) at 3000 ft. are relatively small and show no significant bias to positive or negative values. At all lower heights there is a definite tendency for temperature to decrease during the night; the magnitude of the decrease increases steadily towards the ground and is decidedly greater on "foggy" than on "non-foggy" nights. Humidity changes, on the other hand, cover a wide range at all levels and it is only below 1000 ft. that a systematic

/tendency

tendency to decrease in humidity appears to exceed the random variations. The true systematic change in total water content near the ground is even smaller than is suggested by fig. 7(b), because the diagram is derived from observations of water vapour (dew point), neglecting water present as liquid droplets; the open squares of the diagram should therefore all be moved some distance to the right if changes in total water content are to be indicated.

This irregular behaviour of the humidity profiles is disappointing because the detailed observation of humidity structure was one of the chief novel features of the Cardington work and it had been hoped that some simple pattern would emerge from the results. It seems clear that the greater part of the humidity variation above about 1000 ft. is to be accounted for by the horizontal movement of air with non-uniform humidity; the only alternative is to assume that vertical transport of water vapour occurs to a considerable extent by some mechanism quite unrelated to the humidity gradient and frequently working against the gradient. There is a clear example in fig. 2; the considerable increase in humidity in the region of 1000 ft. from 1800 - 0450 hr. implies either the advection of much wetter air or the vertical transport of water vapour to that region against fairly steep humidity gradients. Similar, though sometimes less clear-cut, arguments can be applied to many of the other large changes in humidity shown in figs. 1 - 6.

It would not be particularly surprising that humidity variations due to horizontal movements should turn out to be greater than those due to vertical movements if the relative simplicity of the temperature profiles did not encourage the belief that advective temperature changes form a small part of the total; the big difference between the temperature and humidity profiles requires some explanation. The humidity fluctuations are far too large to be accounted for by errors in the psychrometer; sampling errors are more serious, as mentioned in Appendix I, but even so the mean deviation of a long series of dew point readings at a fixed height from the smoothed curve drawn through them was only about 0.5°F, corresponding to a change in specific humidity of the order of 0.1 gm./Kg., which would be almost negligible on the scale of fig. 7(b). The most natural explanation lies in the fact that vertical transport of heat can occur by radiation as well as by eddy diffusion, while the only mechanism for transport of water vapour is the eddy diffusion one. We may suppose that as air from different sources, with various temperatures and humidities, moves towards Cardington, the temperature and humidity are modified towards profiles characteristic of the terrain around Cardington. Eddy diffusion usually suffices to modify the profiles up to heights of the order of 500 ft. but above that height the humidity remains characteristic of the original air rather than of the local conditions; the temperature profiles are similarly modified by eddy diffusion but in addition they can be changed by radiative transfer of heat so that they lose evidence of their origin and become representative of local conditions more quickly and to greater heights than the humidity profiles.

A general survey of the profiles thus suggests that it may be useful to discuss the temperature changes up to 3000 ft. in terms of processes going on at one site, i.e. in terms of a simple one-dimensional model, (making allowance for advective effects in individual cases, where necessary) but that it is only in the lowest few hundred feet that humidity changes will conform, even approximately, to such a model. The profiles indicate that 500 ft. is about the height to which the humidity is appreciably affected by local conditions, but it is obvious that there is no sharp boundary between an upper region where advective effects predominate and a lower one where local conditions exercise control. The upper limit for a precise one-dimensional treatment of humidity changes must be well below 500 ft. even for values averaged over many nights and may be lower still when individual occasions are considered and non-uniformity of the immediate surroundings of the instruments becomes important.

If fig. 7 is accepted as showing that humidity changes associated with nocturnal cooling do not usually extend much above 500 ft. it is reasonable

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to infer that the changes in humidity at ground level are not directly affected, to any appreciable extent, by the humidity structure above this level. There will, of course, be individual cases, particularly when a deep fog layer is formed, in which humidity at higher levels is of direct importance, and it may exert indirect effects, by affecting the infra-red emissivity, in other cases.

2.3. Fuller description of typical cases

Despite the difficulties in interpretation caused by advection, it is interesting to see how far the results of figs. 1 - 6 fit a simple qualitative account of the processes at work on a "radiation night".

2.3.1. The simplest type of foggy night (figs. 1 and 2)

(a) Late afternoon - the changing balance

During the afternoon the temperature lapse reaches or slightly exceeds the dry adiabatic value and the specific humidity is constant or decreasing with height (the afternoon ascents would only increase the confusion of figs. 1 - 6 and have been omitted). This corresponds to an excess of incoming over outgoing radiation at the ground, with the balance of heat serving to warm the earth, to evaporate water from it and to heat up the air near the ground. As the sun sinks and the incoming radiation diminishes, the earth's surface cools rapidly to maintain the balance of heat-flow and by 1800 hr. it is several degrees colder than the air above or the soil below. The air near the ground is thus losing heat to the surface both by eddy transfer and by long wave radiation; its fall in temperature is most rapid close to the ground but is noticeable up to 1000 ft. or more. The changes in humidity (as measured by the dew point) are less marked up to 1800 hr. In general there is a small fall in dew point at screen level, presumably because the supply of water by evaporation from the ground is cut off as the heat supply decreases but the upward movement of water vapour by eddy transfer in the direction of lowest specific humidity continues.

(b) Early evening - the end of rapid changes

After sunset the temperature near the ground begins to fall less rapidly, but the dew point, on the whole, falls more rapidly than before. A change in the rate of fall of temperature is to be expected, because before sunset the incoming radiation decreases rapidly and the surface temperature must adjust itself to maintain the balance of heat, but after sunset the incoming radiation (which now contains only a long-wave component) is much smaller and changes much more slowly, so that slow changes of surface temperature suffice to maintain heat balance. (More detailed investigations of this change in rate of fall of temperature, with more elaborate explanations of it, have been reported by Saunders (1949) and by Pedlow (1952); it seems unnecessary to introduce such explanations until the simple mechanism just outlined is shown to be quantitatively inadequate). A change in the rate of fall of dew point is to be expected as soon as the surface temperature falls below the dew point near ground level and dew begins to be deposited; the air near the ground will lose water and its dew point will fall. Fig. 1 and most of the other results shown support the idea that dew begins to form as soon as surface temperature falls below screen level dew point, but fig. 2 shows evaporation continuing even when the surface temperature is recorded as being 7 - 8°F below the screen dew point. "Surface temperature", however, was that measured by a "grass minimum" thermometer with the standard exposure; this is known to give a good approximation to the effective radiating temperature of the surface, which is presumably the temperature of the foliage, but the temperature of the soil surface itself may well be higher, permitting evaporation from the soil even when dew is forming on the grass. This possibility seems to be confirmed by the recent dew measurements of Jennings and Monteith (1954).

/(c)

(c) Early night - the quasi-equilibrium period

After these changes near the time of sunset, the general form of the profiles changes little for some hours. Temperature is low at the ground, rises to a maximum at a height of several hundred feet then falls off again, but at a rate less than the adiabatic lapse rate. There is no abrupt change in temperature gradient at any level and there is therefore little significance in the "height of the inversion"; if this is defined as the height with maximum temperature, it rises, rapidly at first, then more slowly. At sunset the level of most rapid fall of temperature is close to the ground but a few hours later it appears to be higher up, near the level of maximum temperature; the temperature gradient near the ground (below about 100 ft.) thus increases fairly sharply to a maximum soon after sunset then tends to decrease slowly. This is, of course, the sort of behaviour that would be expected to result from the fairly abrupt cutting off of a supply of heat to the surface, with processes such as radiation and eddy transfer acting to spread the cooling upwards at a relatively slow rate. The temperature profiles in the ground are similar, with the vertical scale reversed and much reduced because the thermal diffusivity of the soil is much less than that of the air.

The dew point changes during the same period are, as usual, smaller and more confused by advective effects but the general trend, illustrated in figs. 1 and 2, is for the dew point close to the ground to fall nearly as rapidly as the temperature, while the rate of fall higher up, even as low as 100 ft. on some occasions, is a good deal less rapid. If we consider the difference between temperature and dew point at any level resulting from these changes we see that the first rapid fall of temperature near the ground brings the air there close to saturation, but the subsequent slower cooling and the accompanying fall in dew point following the deposition of dew make further approach to saturation very slow; at higher levels the fall in temperature is later and less abrupt in its beginning, so that the approach to saturation is steadier.

A close association between temperature changes and humidity changes is to be expected if the mechanism responsible for the changes is the upward transfer and mixing of eddies of air cooled by contact with the ground. It has indeed been suggested (Emmons and Montgomery, 1947) that saturation can never be reached by this mechanism alone because water vapour will be transported downwards too rapidly in relation to the heat transfer; this overstates the argument, by neglecting the non-linearity of the temperature-saturation vapour pressure relation, but it is certainly true that eddy transport in nearly saturated conditions will bring about changes in dew point which are, to a first approximation, equal to the temperature changes. Cooling of the air, however, can be caused by radiation (from water vapour and carbon dioxide) as well as by eddy transfer; this type of cooling does not involve any loss of water vapour and can thus help to narrow the gap between temperature and dew point.

(d) The small hours - fog formation

The period with rather slowly changing conditions that has just been considered ends, in figs. 1 and 2, with the formation of fog an hour or two after midnight. Towards the end of the period, in the hours about midnight, conditions at screen level become erratic, with occasional increases of temperature or dew point superimposed on their general slow decrease. These fluctuations are probably connected with small changes in the speed, direction or turbulence of the airflow, but there is no obvious connection with any of the quantities recorded so far. It is very difficult to say what finally determines the time at which fog forms. In both fig. 1 and fig. 2 the screen measurement indicates 100 per cent relative humidity some time before fog is reported. This is particularly noticeable in fig. 2, where there is a period of three or four hours (2100-2400 hr.) during which the screen temperature and dew point are equal and fall by several degrees, yet no general fog forms (though patches of fog were reported in particular directions). It is often suggested that fog formation is brought about by small increases in wind speed and turbulence, but there is little evidence

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for this in the Cardington records. No observations of turbulence were made near the ground, but the pressure-tube anemometer at 140 ft. does not reveal any tendency for turbulence to increase just before fog forms and the sensitive cup anemometers at lower heights show no increase in wind speed. When fog does finally form the detailed visibility records (of which an example is given in fig. 10) show that it usually does so rather abruptly, sometimes in two or more steps. The fact that thick fog often forms quickly (visibility falling from 2500 yd. to 50 yd. in 10 minutes in fig. 10) may be an important clue to the processes involved. Unless it contains a high proportion of very small drops, a fog with visibility 50 yd. must have a liquid water content of at least 0.5 gm./m^3. It seems very unlikely that such a thick fog could be formed by mixing air saturated at different temperatures, as is often suggested, because to produce 0.5 gm./m^3 of liquid water the temperature difference would have to be over 20°F (if the cooler air was at 32°F). On the other hand if 0.5 gm./m^3 is condensed by cooling in a time of the order of 10 minutes, the rate of loss of heat is far greater than the average rate observed before and after fog formation. The simplest explanation (considered in more detail in Appendix V) is that sudden cooling of the air is brought about through loss of heat by radiation from the water drops; it can be shown that as the relative humidity is increased and the size of hygroscopic nuclei grows, a critical size is reached beyond which the loss of heat by radiation from the nuclei leads to unlimited growth - the droplets will stop growing only when they have become large enough to shield each other from further radiative loss.

(e) Late night - the growth of fog

Fog formation is followed by marked changes in the temperature structure near the ground. The screen-level temperature usually shows a fall of a few degrees in the hour covering the period of fog formation but the fall is not always noticeably greater than the fluctuations of temperature normally occurring in the middle-night period; if continuous, short-lag temperature records were available, a closer correlation of temperature change with fog formation might be possible. It is above and below screen level that the most marked changes occur; at the ground surface, the temperature rises sharply to about the screen-level value, presumably as a consequence of the increased back-radiation from the fog, while at greater heights, up to the top of the fog, there is a considerable fall in temperature to values at or below the screen temperature. The result of these changes is to replace the inversion in the lowest layers by an isothermal or lapse region; the fogs of figs. 1 and 2 are too shallow for the precise observed value of the lapse rate to have much significance, but in deeper fogs such as those of figs. 3 and 4 the lapse rate usually seems close to the saturated adiabatic value (3.3°F/1000 ft.). The upper boundary of the lapse region is formed by a steep inversion very similar to the ground-level inversion present before fog formed. The dew point values indicate saturation within the lapse region, sometimes extending a little way into the inversion above, but at greater heights the dew point continues its slow fall, apparently little affected by the formation of fog below, and remains below the air temperature. The few soundings that have been made with the "fog density indicator", combined with visual estimates of fog top height and the results of the humidity measurements, show that the top of the fog is normally close to the level at which the temperature begins to increase steeply with increasing height, i.e. to the base of the inversion.

The fogs of figs. 1 and 2 appear to have reached heights of about 100 ft. shortly after their formation but to have grown very little deeper in the next two or three hours. Fig. 2 shows that after dawn the temperature in the fog layer increased and the height of the fog top rose to at least 250 ft., no doubt because of an increase in turbulence. The growth of fogs can be followed more closely in figs. 3 and 4. Here again a rapid initial growth (much more rapid and extensive than in figs. 1 and 2) is followed, for the rest of the night, by a period of slow change. Fig. 3 shows steady cooling both inside and above the fog layer, with very little change in height of fog top after 2200 hr. Fig. 4 seems to show similar behaviour (though the data are less complete) with the addition of slight warming and upward growth of the fog after dawn.

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A more complete picture of variations in the depths of fog is given in fig. 8. This shows the changes in height of fog top for all occasions on which the height was determined with reasonable certainty, either directly, by the fog density indicator, or indirectly, from the temperature and humidity profiles. It will be seen that there is apparently little tendency for fogs to grow deeper steadily throughout the night; the more typical behaviour is a fairly rapid growth at first, followed by a more or less static period and a final brief period of upward growth as a preliminary to the dispersal of the fog by rising wind and temperature. The sequence of rapid initial change followed by a quasi-equilibrium state is similar to that in figs. 1 and 2 before fog formed, with the difference that in fig. 3 and the other similar cases included in fig. 8 fog formed before conditions had settled down after the change from day to night. The height of the fog top during the quasi-equilibrium period may well be fixed by the need to balance heat exchange there.

(f) Morning - dispersal of fog

The Cardington observations were not planned to study conditions while fog was dispersing. Nevertheless it might be interesting to use them, where possible, for such a study; this has not been done in the present report.

2.3.2. Other types of night

Figs. 1 and 2 were selected for fairly full description and discussion as representing the simplest and most easily explicable type of behaviour. The remaining examples can be discussed more briefly, emphasizing only the points of difference from the description just given.

2.3.2.1. Fog under subsidence inversion

Examination of all the Cardington results gives the impression that there is a second general type of behaviour, less well defined than that represented in figs. 1 and 2, but with certain characteristics worth describing. The important characteristics are that the air temperature falls at roughly the same rate at all heights up to several hundred or even 1000 ft. and that when fog forms it does so in a fairly deep layer, sometimes, in fact, appearing as low cloud which quickly extends down to the surface. These characteristics are most marked when the wind is fairly strong, particularly when it blows from the North Sea (in which case the fog should usually be classed as advection fog not radiation fog) but they are also noticeable in light wind conditions whenever there is an appreciable subsidence inversion in the layer reached by the balloon soundings. Such an inversion usually occurs in prolonged anticyclonic spells so that this type of fog formation is characteristic of the later days of a spell of fog whereas the type of figs. 1 and 2 is characteristic of isolated foggy nights or of the first night of a spell. It is difficult to find an example not complicated by advection effects or by the formation of fog early in the night; fig. 4 is the best that can be found. The effects of subsidence are shown by the warmth and dryness of the air at 1500 ft. and above. It will be seen that the temperature fall (1835-0310 hr.) is about equal at 500 ft. and at screen level, whereas in fig. 2 the fall at 500 ft. was only about half that at screen level.

A possible explanation of the more uniform temperature fall lies in the radiative properties of the air. The air under the subsidence inversion of fig. 4 probably contained fairly large quantities of smoke and other suspended matter (the visibility in mid-afternoon was only three miles for fig. 4 against ten miles for fig. 2) and cooling of the air by radiation from the suspended particles may well have been very appreciable.

The fall in dew point, too, is more uniform with height in fig. 4 than in figs. 1 and 2, and there seems to be a significant loss of water vapour up to nearly 500 ft.; the total amount of water lost from this layer is much greater than in figs. 1 and 2 and is too great to be accounted for by the measured

/deposition

deposition of frost. It seems likely that in this case, with a steep lapse of humidity above 1000 ft., an appreciable amount of water was lost by upward eddy diffusion.

2.3.2.2. Radiation nights without fog

Fig. 5 represents an occasion similar to those of figs. 1 and 2 but one on which fog did not form. There is no important qualitative difference from fig. 2, but presumably a combination of several quantitative factors - the higher initial gap between temperature and dew point, the greater warmth of the ground and the shorter night period available for cooling - were responsible for the fact that the temperature did not fall as low as the dew point and no fog formed.

Fig. 6 is characteristic of several nights with low temperatures when the total amount of cooling of the air (represented by the area between the temperature profiles for 1800 hr. and 0600 hr.) was much less than on the other nights already considered. It is not obvious why the cooling should be so much less; it is true that radiation from a cold ground surface is less than from a warm one, but the back radiation from the atmosphere is also less and the net change in radiative flow is small. The wind was light and the temperature lapse somewhat less than adiabatic, so it seems unlikely that heat was lost by convection to heights above those studied. It is possible that the main reason for the small net heat loss from the air was the liberation of latent heat in the ground as freezing penetrated downwards from the surface. It is impossible to check this hypothesis fully without more detailed measurements in the ground, but it can be shown to be qualitatively plausible (see Appendix III).

The dew point behaviour on this occasion was not significantly different from that on other nights.

2.4. Wind structure

The balloon can only be used when the surface wind is less than 15 kt. and the wind aloft less than about 30 kt. There are thus no observations for occasions of strong winds. However, the fact that observations were made only when fog or radiation conditions were forecast would in any case have excluded occasions with strong winds; there are results for roughly equal numbers of occasions with winds of 0 - 10 and 10 - 20 kt. (at 2-3000 ft.), but for many fewer occasions with winds over 20 kt. The (2000 ft.) wind speed, below about 20 kt., appears to have little effect on the general course of events; there is no obvious difference between the group of results with speeds 0-10 kt. and that with speeds 10-20 kt. The few cases with speeds over 20 kt., however, were all ones when fog did not form, apparently because the fall of temperature near the ground (though not the mean fall in the lowest 1000 ft.) was much less than usual, as would be expected if turbulent mixing was occurring.

The wind profiles shown in figs. 1-6 are typical of all the results obtained. It will be seen that individual points are somewhat erratic, probably because the wind was measured over only fairly short time intervals ($\sim$1 min.) at each height. Mean profiles for 1800 hr. on ten nights which remained clear and ten which subsequently became foggy are shown in fig. 9. The most interesting feature is the decrease in speed with increasing height above about 750 ft. on fog nights. Even in the mean profiles this decrease amounts to some 30 per cent of the maximum speed; in the individual profiles (e.g. 2200 hr. in fig. 1) it is often much greater, and occurs with great regularity, particularly in the period just before fog forms. Indeed, examining the last ascent before fog formed in each case, we find that on 19 out of 21 occasions with complete enough wind data the wind at 3000 ft. was less than that at some level below 2000 ft. On the other hand, on non-foggy nights this decrease with increasing height occurred on only about half the occasions, and when it occurred was less marked than on foggy nights.

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The simplest explanation of the decrease in wind speed with increasing height is that it is a "thermal wind" effect. A decrease should occur if there is a horizontal temperature gradient in the opposite direction to the pressure gradient. Such conditions should exist in the anticyclonic situations typical of foggy weather, either because the air temperature is lowest near the high pressure centre or because the subsidence inversion is highest there, giving the greatest depth of cold air. The winds are so light that quite small gradients would produce the observed effects - gradients of the order of 2°F in 100 mi., equivalent to a 10°F inversion with a slope of 1 in 500 - and the observations available are not exact enough to test the explanation just given.

Below about 1000 ft. the mean profiles of fig. 9 show the effect of friction at the ground in reducing the wind speed near the surface, but it seems clear that the layer in which this friction is the chief factor determining the profiles is only 500 or at most 1000 ft. deep. The wind profiles thus resemble the humidity profiles rather than the temperature ones and tend to confirm the idea that so far as eddy diffusion is concerned the effects of the surface do not usually extend much above 500 ft.; the changes in temperature at greater heights can be attributed to radiative effects (or, possibly, to a greater effectiveness of eddies in transporting heat).

The changes in wind speed during the night show little regularity, though, as can be seen in figs. 1 - 6, they may be large. Close to the ground the wind usually drops in the late afternoon, but after 1800 hr. there is little systematic change until fog forms, when the speed often drops below the range of the anemometers (1 ft./sec.). This calm often extends through the whole depth of the fog, so that there may be a very noticeable drop in wind speed at levels of a few hundred feet when the fog reaches the level concerned. At greater heights large changes in wind speed are common, with decreases on the whole more common than increases. This may well be because rising winds are often associated with approaching fronts and cloudy conditions, whereas falling winds are common in ridges of high pressure, with their clear skies. The practice of making observations only on clear nights (or ones becoming foggy later) thus introduces a bias against occasions with increasing winds.

The wind profiles seem to be rather little affected by the vertical temperature structure. In fig. 2, for example, the wind profile changes little in the lowest 1000 ft. despite the growth and intensification of a considerable temperature inversion.

2.5. Horizontal and oblique visibility in fog

We have so far discussed fog in terms of its presence or absence, without mentioning its intensity, which is conveniently measured by the "visibility" in the fog. It will be seen that in all the examples given there was, in fact, a rather abrupt change from a state of no fog, with visibility of the order of 1500 yd., to one of thick fog, with visibility between 50 and 200 yd. Examination of all the records from Cardington shows that visibilities between about 200 and 1000 yd. are rare, except when fog forms at temperatures well below freezing point. This suggests that most thin fogs are in some way unstable; it is possible that this instability is produced by radiative cooling of the fog droplets, in the way discussed in Appendix V.

The final values reached by the visibility all lie within a rather narrow range (50-200 yd. except for very cold fogs) and the visibility usually remains fairly constant for most of the life of the fog. It might be thought that the visibility should be related to the extent to which the temperature falls below the temperature at which the air becomes saturated, but no such relation can be seen in the Cardington data; in particular it can be seen in fig. 3 that the temperature throughout the depth of the fog fell by some 3°F between 2030 and 0730 hr. while the visibility fluctuated about a constant value of 50 yd.

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The variation of visibility with height is of considerable importance to aircraft operators. Measurements to investigate this have been made on a few occasions since 1953 and the results are shown in fig. 11. The quantity measured and plotted is the one of direct interest to aircraft, the slant visual range of a lamp at various heights to an observer at ground level; it represents the integrated effect of the fog in the layers between the observer and the lamp and therefore varies much less abruptly with height than the fog density itself. Readings from the "fog density indicator" agree qualitatively with the slant visibility observations but are thought to be less accurate because of difficulties in calibration, and are not used in fig. 11. It will be seen that the observations are too few to allow general conclusions to be drawn (they were nearly all made in one foggy spell) but their indication is that slant visibility commonly decreases with increasing height, the decrease being most marked in relatively thin fogs. A theoretical interpretation of these results is given in Appendix VI.

2.6. Summary of qualitative results

General study of the Cardington results, as indicated in the last few sections, shows that there is little difficulty in giving a satisfactory qualitative account of events on a foggy night in terms of physical processes. For a fuller understanding of these processes we must turn to a more exact quantitative treatment of the data. Before doing so, it is well to review the main features of the general picture we have outlined. The dominating process is obviously the loss of heat by radiation from the ground and the balancing of this loss by downward radiation, by heat convected from the air or conducted through the soil and by latent heat released near the surface. Transport of water vapour is also important, but there is good evidence that in many cases only transport in the lowest 500-1000 ft. need be considered and that this transport is largely balanced, as far as the approach to saturation is concerned, by the simultaneous transport of heat. It appears that large-scale advective changes of temperature are usually small compared with the time-change during a night, so that one-dimensional treatment of heat flow may be practicable, but that advective humidity changes are much larger than the time-changes, except quite close to the ground, and make a one-dimensional treatment of very doubtful validity. The wind structure appears to affect and be affected by the temperature structure rather less than might be expected, which may permit some simplification in theoretical treatments. For the formation of fog at Cardington, the attainment of a relative humidity of at least 99 per cent is a necessary condition for fog formation (at least, at temperatures above freezing point) but it is not a sufficient condition. Certain additional conditions must, apparently, be fulfilled before fog forms, but the present data are insufficient to decide what they are; they may well be connected with the detailed humidity structure and radiative properties of the atmosphere over the site, but small horizontal movements of inhomogeneous air may also be essential.

Before attempting a more quantitative examination of the processes discussed, it is worth pointing out the few features of the general picture of events that might be clarified if a longer series of observations were available (and assuming that the complete plan of observations set out in 1.1. had been carried out). The most important, and probably the most complicated, is the effect of wind and cloud on the cooling process. Observations have so far been deliberately restricted to more or less cloudless nights and most of them have been made in relatively calm conditions. The effect of wind and cloud are already known approximately, but it is possible that observations on windy and cloudy nights would bring some new facts to light. Another subject on which more data might be valuable is the vertical extent and growth of fog, particularly when combined with information about the visibility (i.e. the density of the fog at various levels). The present data do give a general idea of typical behaviour at Cardington, but are not extensive enough to give accurate averages, which may well be required in aircraft or atmospheric pollution problems.

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3. Quantitative discussion of physical processes

Although the physics of the various component processes leading to fog formation is fairly well understood and it is possible to calculate the results of each process in simple conditions, the interactions between the different processes lead to such complexities that there has so far been little success in calculating the important features of the final state on a foggy night from observable initial conditions. The first step in disentangling the problem is to find out the relative importance of the various processes at work; we can then try to estimate whether or not the current theory for each process fits the facts with sufficient accuracy, and seek to make improvements in the theory or empirical approximations to the required results where necessary.

3.1. Heat balance

It is clear from fig. 7 and the temperature profiles shown in figs. 1 - 6 that the main temperature changes on a radiation night are usually confined to the lowest 3000 ft. or less. The heat balance of this "cooling layer" has the heat lost by radiation through the top of the layer on one side and on the other side the heat given up by cooling of the air, by cooling of the earth and by condensation or freezing of water. In addition we may have to allow for heat flowing through the top of the cooling layer by convection and for heat transferred horizontally by advection. The components of the heat balance on a number of occasions already discussed (figs. 1 - 6) are set out in Table I. The methods of finding the various components from the observations are indicated below.

Table I

Rate of loss of heat from cooling layer. Units mW/cm^2

Date	29/9/51	29/9/51	27/9/53	5/12/52	31/10/53	18 foggy nights	13 clear nights
Time	0001	0400	0100	0001	0001	18 - 06	18 - 06
Level of top of cooling layer	2000ft.	3000ft.	3000ft.	1000ft.	1500ft.	2500ft.	2000ft.
Observed loss of heat							
(a) from air	5.0	11.7	7.8	1.8	1.6	5.9	3.0
(b) from earth	5.0	4.5	3.9	1.0	3.6	4.5	3.0
(c) from condensing water vapour	0.7	0.7	0.4	0.4	0.7	1.0	0.5
Total observed loss	10.7	16.9	12.1	3.2	5.9	11.4	6.5
Loss by radiation through top of cooling layer (calculated from "Kew chart")	9.3	9.0	9.5	7.6	8.8	9.0	8.2

In all columns there is a large difference between the figures in the last two rows, indicating a serious lack of balance between the components considered. The most uncertain of the components is the heat lost by the earth. This was estimated from temperature profiles drawn through earth thermometer readings at 2 in., 4 in. and 8 in. in the ground, by taking

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the area between profiles at two-hour intervals and multiplying by a mean capacity of 0.7 cal.cm^{-3} °C^{-1} measured at Cardington some years ago by Swinbank. The temperature profiles are not determined very closely by the three temperature readings and the heat capacity may well have varied with the condition of the soil, so that the figures in Table I may be in error by as much as 50 per cent. On 5/12/52 and 31/10/53 there is another source of error in the earth-heat estimates because the surface temperature was below freezing point and any water present in the soil was presumably freezing and giving up latent heat. The freezing of 1 mm. of water per hour (i.e. of 2-3 mm. of damp soil per hour) would release about 8mW/cm^2 of heat, so this latent heat component can easily be an important one.

It is thus possible to explain the discrepancies in all but the second case of Table I in terms of errors in the estimates of the heat contribution from the earth. There may, of course, be other errors in the table; the loss of heat from the air and the loss by radiation are probably correct to 10 per cent but the latent heat of condensation of water depends on the measurement of dewfall, which is only accurate to about 30 per cent. As far as convective transport of heat through the top of the "cooling layer" is concerned, we may note that in all cases the lapse rate was less than the dry adiabatic one, so that convection would convey heat downwards and would thus increase the discrepancy in the first three cases of Table I but reduce it in the fourth and fifth. In the fourth case the wind speed was low and convection through the 1000 ft. level probably small, but in the fifth case the wind was fairly strong and it seems better to attribute the unbalance shown in Table I to the downward transport of heat than to latent heat of freezing in the earth (the surface temperature fell only 3°F below freezing point).

In the second case of Table I none of the errors so far invoked can explain the discrepancy, and we are forced to attribute it to an advective change in air temperature.

The last two columns of the table give averaged estimates of the heat balance components for 18 foggy nights and 13 clear ones (the nights for which data were given in figs. 7(a) and 7(b)). The air and earth losses should have the same accuracy as in the individual cases of Table I, but the condensation and radiation components are rougher estimates, because the necessary measurements and calculations were not made on all occasions. The heat of condensation for the foggy occasions assumes that, in addition to an average amount of dew, water condensed to form the fog to an extent of 0.5 gm./m^3 in a layer 200 metres deep.

There is a striking difference between corresponding heat components on foggy and non-foggy nights. The explanation probably lies in the high proportion of the non-foggy nights which had freezing ground, and release of latent heat (as on 5/12/52), or strong winds, and downward transport of heat from above (as on 31/10/53).

These results in Table I are very similar to the longer series obtained by Swinbank (1946) from earlier observations at Cardington. Both sets of results show that the outward flux of radiation (as calculated from a radiation chart) is relatively constant from night to night but that the observed losses of heat from air and earth (and also their sum) are considerably more variable. Some of this variability may be due to errors in measurement, but advective changes also seem to play a significant part and it is quite possible that the "chart" estimates of radiation may be inaccurate on some occasions. Swinbank, indeed, suggested that his results demonstrated a systematic error in the radiation chart (Elsasser's) he used, but the general scatter of his results (as of the present ones) appears too great to justify such a conclusion. On the other hand, it cannot be said that the results provide any serious check on the accuracy of radiation charts - this is done far better by direct measurements of radiation.

3.2. Heat exchange within the cooling layer

Though the study of the heat balance of the cooling layer as a whole casts doubt on the accuracy with which the various components can be estimated from the Cardington data, it is interesting to carry the analysis a stage further and try to find the components of heat flow at a series of levels within the cooling layer. This has been done for a few cases, and results are given in Table II. We again start by finding the flux of radiation, using the observed temperatures and humidities in conjunction with a radiation chart. The first three columns of Table II show the upward, downward and net fluxes calculated in this way for various levels, including the ground surface. The difference in net flux at any two levels (col. 4) gives the loss of heat by radiation from the intervening air-layer. The observed loss of heat from the layer (col. 5) can be found by multiplying its heat capacity by its mean rate of change of temperature, and the difference between this loss and the radiative loss measures the loss of heat in the layer by convection (col. 6). Col. 6 thus gives the difference in convective flow between each pair of levels; in order to find the actual magnitude of the flow, some boundary condition is needed. This could be provided by the balance condition at the earth's surface, if all the other components there were known, but we have seen that the value of heat flux from the earth is not known precisely from the measurements, and it has seemed better to assume the condition that the heat flow through the top of the cooling layer was zero (in the first three cases analysed in Table II; in the last case, as noted in discussing Table I, this assumption seems unreasonable and the alternative boundary condition provided by the balance equation at the earth's surface has been used). With these conditions, we can use col. 6 of Table I as a set of differences to build up col. 7, giving the convective flow through each level. To complete the list of heat flow components at the surface, the latent heat of condensation of dew (the product of the latent heat and the measured dewfall) and the heat lost by the earth are given in cols. 8 and 9. For balance, the sum of the ground-level values in cols. 7, 8 and 9 should be equal to the value of net outward radiation in col. 3. The obvious lack of balance has already been discussed in connection with Table I and attributed mainly to errors in heat flow from the earth (col. 9) - but the possibility of systematic error in the convection components (col. 7) must also be borne in mind.

The first interesting feature of Table II is the broad equality between cols. 4 and 6, the radiative and convective cooling of the air, within the "cooling layer".

The main analysis into radiation and convection components in Table II is independent of errors in the heat flow from the earth, but, as will be obvious from the table, it is very sensitive to errors in the estimation of the upward and downward fluxes of radiation, particularly when shallow air layers are involved. It is shown in Appendix II how the usual process of estimation from a radiation chart can be rearranged to give changes in net flux directly (i.e. col. 5 of Table II, rather than cols. 1 and 2) and thus allow greater precision, provided the observations and the data on the radiative properties of the atmosphere are accurate enough. It is easy to show, however, that the radiative properties of a hazy or misty atmosphere are very different from those of the clear air to which radiation charts refer, and that quite small amounts of liquid or solid particles can seriously upset the sort of calculations on which Table II is based.

Even when radiation from particles is ignored, as in Table II, we find that radiation is, broadly speaking, as important as convection in cooling the air. Radiative cooling is seen to be greatest near the ground but to fall off rather slowly with increasing height; even at 2-3000 ft. it is usually of the order of 1 mW/cm^2 per 1000 ft. layer, equivalent to a rate of fall of temperature of about $0.2^{\circ}F/hr$. In most cases, as we have seen, no appreciable fall in temperature is observed above about 2500 ft. and we must assume that the air at these heights usually gains heat in some way to compensate for the loss by radiation. It seems likely that this heat is frequently provided by anticyclonic

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As far as can be seen from Table II, convective cooling is usually a maximum at heights of a few hundred feet; it certainly falls off at greater heights, but the analysis is not exact enough to be certain of the behaviour near the ground. It is likely that the radiative cooling becomes very large (and the convective component correspondingly small) at heights of a few feet, while at heights of a few inches where the air is close to the surface temperature of the ground, with considerably warmer air above, radiation must warm rather than cool the air. The heat economy of these low levels is interesting but cannot be studied with the present observations.

It can be seen that the convective flux normally changes by an amount of the order of 10 per cent in the lowest 100 ft. setting a limit to the height to which the assumption of approximately non-divergent flux is valid. The convective fluxes in col. 7 of Table II can be combined with the observed gradients of potential temperature (col. 10) to obtain values for the "eddy conductivity", K_H, given in col. 11. (The values given at heights of 0 ft. were calculated from the flux at 0 ft. and the gradient at 50 ft.; they thus represent the eddy conductivity at 50 ft., assuming that the flux there is the same as that at the surface). The zero values of K_H at certain heights are, of course, a direct consequence of the assumption that the convective flux there was zero; in view of the smallness of the temperature gradients at these heights a quite small flux would correspond to a large value of K_H, and little reliance can be placed on the values in the table for heights of 1000 ft. and over. At lower heights the values of K_H should be correct as to order of magnitude but are obviously by no means exact. It would be unjustifiable to use the figures to deduce a law for the variation of K_H with height, but there is some indication that changes in K_H in the layers between 50 and 500 ft. are small compared with changes in temperature gradient. To illustrate the relation between the observed values of K_H and the stability, values of the Richardson number for the 50 ft. level are given in col. 12 of the table and values of K calculated from Rossby and Montgomery's formula ($K = \kappa^2 z^2 (\partial u/\partial z)(1+\sigma Ri)^{-1}$ with $\kappa = 0.4$, z = height, u = velocity, σ = const. = 8, Ri = Richardson number) in col. 13. It will be noticed that the values of Richardson number are a good deal higher than in most previous studies of eddy diffusion, partly because of the great stability, but partly because they refer to a relatively great height. There appears to be very little connection between the values of K_H given in col. 11 and either the Richardson number or the calculated values in col. 13, but it must be pointed out that the gradients of temperature and wind speed, on which these calculations depend, are not at all well defined by the observations.

Table II is interesting as indicating orders of magnitude of certain quantities on typical radiation nights but is clearly unsuccessful as an attempt to apply to the Cardington data the type of analysis which has been used successfully in detailed investigations of the lowest few metres. Unfortunately it does not seem possible to say whether the failure is due to lack of precise observations or to the more fundamental reason that a one-dimensional treatment of events is inadequate.

3.3. Water exchanges

We can attempt to build up a table indicating the movement of water in the way Table II indicated the movement of heat. The process is simpler, because radiation plays no part; the basic data are the flux of water at the surface (the dewfall) and the changes in water content of each layer of air. Satisfactory dew measurements were made on relatively few occasions and Table III includes all cases for which an analysis can be made. (Water transport after fog formed cannot be studied because no measurements were made of the water content of the fogs).

It will be seen from Table III that although the measured amounts of dewfall are of the same order of magnitude as the changes in water content of the

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TABLE III

Analysis of flow of water vapour on selected occasions

Units: micrograms / cm^2/sec. (gradients in μ gm./gm./cm.)

Date		18–19/9/53	26–27/9/53	30–31/10/53	28/2/53	1/3/53	6–7/3/53	8–9/3/53	10–11/3/53	Mean	Mean deviation
Period		2108–0310	2155–0100	1800–0001	1800–2400	1800–2100	1800–0600	1800–0600	1800–0200		
Measured dewfall		0.28	0(<.05)	0.30	0	0.20	0.18	-0.02	0.17	0.14	± 0.11
Change in water-vapour content of layer	0–100' 100–250' 250–1000'	-0.20 -0.18 -0.95	0.32 -0.32 +0.63	-0.02 0.03 +0.29	-0.01 0 -0.10	-0.15 0.30 -0.32	-0.05 -0.08 +0.10	0.04 +0.05 +0.09	+0.07 +0.10 +0.04		
Flux of water (+ = downwards)	0 100' 250' 1000'	0.28 0.08 -0.10 -1.05	0 -0.32 -0.64 -0.01	0.30 0.28 0.25 0.54	0 -0.01 -0.01 -0.11	0.20 0.05 -0.25 -0.57	0.18 0.13 0.05 0.15	-0.02 -0.06 -0.01 +0.08	0.17 0.24 0.34 0.38	0.14 +0.05 -0.05 -0.07	± 0.11 ± 0.13 ± 0.21 ± 0.38
Specific humidity gradient (+ = incr. upwards)	0 100' 250' 1000'	- 0.13 0.002 0.006	- 0.23 0.08 0.006	- 0.092 0.033 -0.015	-0.01	-	0.023	-0.09	<0.02		

lowest few hundred feet of air, there is very little correlation between the water lost from the air on any occasion and that deposited on the earth. The table includes estimates of the gradients of specific humidity at each level. These are not very accurate, because of the irregularity of the profiles, but they do show clearly that there is practically no correlation between the gradient and the flow through any level; in fact, in six cases out of twelve the flow is apparently against the gradient. This result seems to indicate that advective changes are swamping the effects of vertical transport even at heights as low as 100 ft. The "advection" involved is not necessarily on a large scale; it may be that the horizontal movement of quite small patches ($\sim$ 100 yd.) of damp or dry air makes the "spot" humidity readings on which Table III is based unrepresentative of mean conditions over the site.

Although Table III suggests that water diffusion coefficients calculated from the Cardington data will have little significance, there is some interest in comparing them with the values of K_H in Table II. This can be done in only two cases, 27/9/53 and 31/10/53. Following the procedure used in Table II we estimate the diffusion coefficient, K_W, at 50 ft. by using the flux at the surface (dewfall) and the humidity gradient at 50 ft. In this way we find, for 0100 hr. on 27/9/53, K_W = 200 cm^2/sec. (K_H was 2,000 cm^2/sec.) and for 0001 hr. on 31/10/53, K_W = 1700 cm^2/sec. (K_H was 2,800 cm^2/sec.). Errors in the measurement of dew and of humidity gradient can easily account for the discrepancy between K_H and K_W in the second case; in the first case the discrepancy suggests that K_H has been overestimated as well as K_W underestimated, or else that there is a genuine difference between K_H and K_W, defined in the way implied by the method used for calculating them.

We can attempt to overcome the irregularity of the humidity results on individual occasions by taking averages. Table III therefore includes mean values (with mean deviations) for the flow through each level. There appears to be a significant downward flow at the surface and at 100 ft., but at 250 ft. and even more at 1000 ft. the spread of individual values is so wide that the mean flow cannot be determined accurately. The most that can be said is that the results do not contradict the simple hypothesis that the mean flow several hundred feet above the surface is zero, i.e. that the normal loss of water from the lower layers on a radiation night can be wholly attributed to the deposition of dew.

3.4. Measurements required for the investigation of physical processes

The last few sections have shown the order of magnitude of the various terms in the heat and water budgets on a radiation night. It turns out that the three most obvious components of heat flow at the earth's surface - radiation, conduction from the earth and convection - are of roughly equal importance. Slightly less obvious results are that direct radiation from the air contributes at least as much to its cooling as does convection and that latent heat of freezing of water in the earth can provide a significant amount of heat on some occasions. As far as water is concerned, the outstanding result is the frequent occurrence of large changes in water content which, since they are apparently uncorrelated with vertical humidity gradient, must be attributed to horizontal advection. On average, the amount of dew deposited is comparable with the amount of water lost from the air.

More detailed examination of the results given in Tables I - II showed that they could not be used to check or develop theories of the various processes going on. On radiation no measurements at all were made, on conduction in the earth the measurements were too incomplete and for convective processes the measurements of gradients showed considerable observational uncertainty, while the estimates of flux were very sensitive to small errors in the radiation or conduction terms. It is important to try to decide what accuracy is needed in any future measurements if they are to contribute

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effectively to our understanding of the physical processes of fog formation. This involves rough theoretical treatments of various processes, which are set out in Appendices II - IV. Only the main results will be mentioned here.

3.4.1. Radiation

It is shown in Appendix II that for calculating heat balance at the surface an accuracy of the order of 2 per cent is required in estimating the "back radiation" from the sky. This is about the limit of accuracy obtainable from a radiation chart (Robinson, 1947) provided the estimate is not upset by the "additional radiation" tentatively attributed by Robinson to haze particles. It seems essential to find by direct measurement how large this additional radiation can become on a hazy night. When we come to calculate the direct radiative cooling of the air, the effects of haze seem even more important; Appendix II suggests that if the visibility is as low as 2500 yd. cooling by radiation from the haze or mist may equal that by radiation from water vapour. Direct measurement to test this conclusion seems very desirable. The accuracy of measurement required for both back radiation and radiative cooling measurements is probably just within the capacity of ordinary radiometers, if used with care.

Another point which requires attention is the variability of ground surface temperature from place to place. "Surface temperature" in the Cardington work has been given by the readings of a thermometer resting just above the ground, as recommended by Robinson (1947). This is probably representative of the immediate surroundings of the site, but if we are concerned with the equilibrium of a 3000 ft. layer of air, variations over a wider area, including hedges and ploughland as well as grass, may be important. It might be possible to measure a mean radiating temperature by a downward-directed radiometer from a balloon.

3.4.2. Heat from the ground

For heat balance calculations such as those giving Table I and for acceptable accuracy in predicting the course of temperature changes it is probably sufficient to know the heat flow from the earth to about 10 per cent. This is impossible with the present earth temperature measurements and without measurements of the specific heat of the soil, but it has always been planned to increase the number of depths at which earth temperatures are measured and to make specific heat determinations, and these extra observations should give the required accuracy. The use of "flux plates" as developed at Kew, might simplify the measurements.

As far as the theory of heat transfer in the earth is concerned, there is little to suggest that ordinary heat conduction theory is inadequate, though when more detailed temperature profiles are available it is possible that difficulties will be revealed. As with radiation, the question of the representativeness of the site and the possible effects of variation in heat flow from place to place must be borne in mind.

Complications in the calculation of heat flow occur when the ground is freezing, as considered in Appendix III. In order to measure the heat flow from the ground, special measurements would be necessary in these conditions.

3.4.3. Convection processes

Convection is the most difficult transfer process to deal with, as there is no fully worked out theoretical treatment available and no generally accepted method of measuring it directly. In Table II we

tried to estimate the convective heat flow by the usual indirect method, but found that the data on which we relied were not accurate enough to give good estimates. If measurements of radiation and more accurate measurements of earth flux were made, as just suggested, reasonably accurate estimates of convective flux should become possible, so long as advective effects are not too large. There is evidence that they will probably be large enough to upset the results on a fairly large proportion of occasions, even when synoptic charts show no obvious signs of advection.

Given satisfactory estimates of convective flux, we shall require to relate them to the gradients of wind and temperature. The profiles obtained by the present procedure are too erratic to give good estimates of gradient, as can be seen from the examples in figs. 1 - 6. This is not surprising since, even in work in the lowest few metres, it is usually found necessary to average over periods of the order of half an hour in order to get smooth profiles, and it is unlikely that any much shorter time of averaging will suffice at greater heights.

The most that can reasonably be hoped for from a study of convection processes is that it may be possible to find empirical expressions relating the eddy diffusion coefficients to wind and stability parameters. This has been done (e.g. Rider, 1954) in near-neutral conditions of stability and for the lowest layers of the atmosphere, where the divergence of the convective flux is negligible. To extend this type of treatment to extremely stable conditions and to deep layers where the product of diffusion coefficient and gradient is known to vary with height is clearly a big undertaking. There would be little hope of predicting fog formation by physical methods if the result depended on calculating from a convection theory the rate of cooling of air at a given height or the loss of water vapour from the air. The chief reason for thinking that some sort of success may be possible is that we are interested in obtaining saturated air, and with nearly saturated air the rates of change of temperature and of dew point by convection should be nearly equal, provided the eddy diffusion coefficients for heat and for water vapour are the same. The question of the combined eddy diffusion of heat and water is considered in Appendix IV and it is shown that a relatively rough value of the diffusion coefficient and its variation with height is all that is required, so long as the coefficient is the same for heat and for water. The question of the equality or otherwise of "K_H" and "K_w" is thus one of great importance. It is suggested that the first aim of future experiments should be to settle this question for the conditions of a radiation night. The very erratic nature of dew point variations above about 100 ft. suggests that the experiments will have to be confined to the lower layers, in the first place, if they are to have any chance of success.

3.4.4. Fog formation

It was mentioned in section 2.6 that there is some doubt about the conditions necessary for the actual formation of fog. It seems impossible that fog could form as suddenly as it often does by the mere continuing of the cooling processes considered in the last few sections. Two other mechanisms can be suggested, the effects of horizontal air movements in mixing air at different temperatures and the effects of radiative cooling in promoting the growth of fog drops once they have formed. The Cardington observations do not allow any detailed investigation of either mechanism, but a preliminary theoretical examination of the radiation mechanism is made in Appendix V and it is concluded that it may well be important. Experimental study of these questions requires measurements of radiation while fog is forming and a study of local temperature and humidity fluctuations.

3.4.5. Fog density and structure

The density or "visibility" of a fog may vary widely, as may its height and the way in which density varies with height. The fog problem will not

be solved until all these quantities can be predicted, but the observations made so far do not suggest any obvious ways of doing so. It seems unlikely that much progress will be made on the experimental side without measurements of the size and number of the water drops of the fog and possibly also of the nuclei around which they are formed. On the theoretical side studies such as those of Best (1953) on the growth of water drops should lead to useful results, though the radiation factor mentioned in the last section should probably be taken into account. For the variation of fog density with height (measurements of which were mentioned in 2.5) a provisional explanation is given in Appendix VI, but it cannot be tested quantitatively without measurements on the water drops.

4. Empirical fog-forecasting methods

In view of the difficulties of a full physical treatment of the processes leading to fog formation, one of the most attractive approaches to the practical problems of forecasting fog is by constructing semi-empirical formulae or graphical methods. The method is to choose certain parameters which seem likely to be important and arrange them in some plausible formula whose coefficients can then be obtained by appeal to experience through statistical or graphical methods. Forecasting procedures of this type have been suggested, among others, by Briggs (1950), Craddock and Pritchard (1951), Saunders (1950, 1954) and Swinbank (1949). The Cardington data can be used to test these procedures and to study possible extensions of them.

4.1. Tests of suggested methods

Briggs, Craddock and Pritchard, and Saunders all suggest methods for predicting the "fog-point", i.e. the temperature at which fog is expected to form. Fog points (T_f) have been calculated by each of these methods for every suitable night at Cardington, and compared with the actual temperature at which fog formed. (This work was done by Mr. D.B.B. Powell). The results are shown in figs. 12 - 14. It will be seen that the forecasting errors are large and are about the same, on average, with each of the methods (in each case the standard deviation of forecast minus actual fog points is about $5^{\circ}F$).

Swinbank's "fog prediction diagrams" give a different form of result. Points representing the initial conditions are plotted on the diagram (fig. 15) and it is found that lines can be drawn separating the points corresponding to fog formation from those corresponding to clear nights; the diagram can thus be used to predict whether fog will form or not, but not to predict the actual temperature at which it will form. The Cardington results are plotted on a diagram of this type in fig. 15. The diagram is rather confusing to interpret but the results are summarized in Table IV.

Table IV

Number of cases

	Predicted		
Actual	Widespread fog	Patchy or local fog	No fog
Widespread fog	6	8	1
Patchy or local fog	3	4	2
No fog	1	5	0

/Figs. 12 - 15

Figs. 12 - 15 suggest that the four methods tested are all rather inaccurate, but it should be pointed out that the data have been treated in a mechanical manner, without making allowance for special effects such as those of wet ground, frost or advection (though cases with obvious changes of air mass have, of course, been excluded) as recommended by the various authors. In addition, the methods have been applied to a different place from the ones for which they were designed (and in the case of Swinbank's method, which was intended for use in October and November only, to other times as well). In view of the small number of cases for which data were available, it seems impossible to allow for these effects without risking a tendency to force the observations to fit the theory in mind. For objective statistical testing of the empirical type of forecasting formula it seems essential to have of the order of a hundred cases, rather than the twenty so far available from Cardington.

4.2. Development of forecasting methods

The simplest of the forecasting formulae considered is that of Craddock and Pritchard, whose initial parameters are simply the screen level temperature and dew point (with an allowance for cloud amount and wind speed when these are not small). It is interesting that this formula seems about as successful as the others, which introduce additional, upper air parameters. As suggested by Craddock and Pritchard, their formula provides a useful starting point for the study of the importance of other parameters. It has been used in examining the Cardington results, by finding, for each case, the error (forecast - actual T_f) of the Craddock-Pritchard formula and plotting this error against various parameters - "hydrolapse" (average vertical gradient of specific humidity in the lowest 1000 ft.), wind shear and earth temperatures. No correlation was found with any of these quantities; this is surprising, in view of the importance given to hydrolapse in other forecasting formulae, but it may well be that a genuine effect of hydrolapse is being masked by large random variations due to other parameters, in the small sample of occasions from Cardington. It seems unlikely from these results that neglect of any one factor is responsible for the "errors" in current forecasting formulae. If there are several independent sources of error, the need for a reasonably large sample of occasions for study is clear.

A less systematic study has been made of various other parameters - humidity and temperature gradients at different levels, wind shear in various layers and heights of inversion levels - but no significant results have been obtained. The only interesting relationship found is one between the formation of fog and a parameter not observable at the beginning of a night, the "grass minimum" temperature. The basis for this relation was pointed out many years ago by Taylor (1917); if the air above the ground is cooled by mixing with air actually at the surface (which is assumed to have temperature and dew point equal to the surface temperature) the mixture cannot be saturated unless the surface temperature is lower than a certain value, determined by the temperature and humidity of the air above the surface and the curvature of the temperature - saturation vapour pressure relation. This suggests that, with given initial (screen level) temperature and dew point, fog cannot form unless the surface ("grass minimum") temperature falls below a certain, calculable value. We have seen that cooling of the air by radiation is also important and will upset this argument, but we have also seen (Appendix II) that radiative cooling near the ground is largely controlled by the difference between air and ground surface temperatures, so that it is possible that an empirical modification to the calculated surface temperature required for fog formation may allow for the radiation effect.

Fig. 16 illustrates a search for this hypothetical relationship. The initial temperature (T) and dew point (D) are represented by co-ordinates in the plane of the diagram and the observed grass minimum temperatures (T_m) by the co-ordinate perpendicular to this plane. (The values of T_m, in degrees Fahrenheit, are written at the appropriate places in the diagram). If the hypothetical relation between T, D, T_m and fog formation exists, there must

/be

be a surface (which can be represented by contours of T_m) separating the points of fig. 16 which correspond to foggy nights from those which correspond to clear nights. The T_m figures in fig. 16 corresponding to clear nights have been ringed and it will be seen that contours can, in fact, be drawn so that all ringed values of T_m are above the corresponding contour values and all unringed values below (with three or four exceptions in twenty-four cases). The 35°F contour to be expected on Taylor's theory (i.e. with radiation playing no direct part in cooling the air) is shown as a broken line; as might be expected, it has the same trend as the empirical contours, but is displaced upwards.

Fig. 16 thus appears to afford a fairly satisfactory way of forecasting fog at Cardington, provided the grass minimum temperature can be forecast to within a degree or two. This is, of course, a useful datum to take because it summarizes the effects of cloud, wind, stability and earth temperature, but the problem of predicting it is correspondingly hard.

5. General conclusions; future programme

5.1. Results of the investigation

The Cardington results have been discussed under three headings. Under the first it was shown that the results provided a fairly detailed general picture of events on a foggy night, which could readily be given a qualitative interpretation in terms of physical processes. The two most unexpected features of the general picture are the great variability of dew point (compared with that of temperature) at heights between, say, 300 and 3000 ft. and the form of the wind profiles, which almost always show a maximum speed in the layer 1 - 2000 ft. with lighter winds above, in the period before fog forms. It seems unlikely that much will be added to the general picture by continuing the present programme of observations at Cardington. Even the two unexpected features mentioned above could best be studied, in the first place, by observations at some other site, to make certain that they are not due to local peculiarities. It is suggested, then, that no further observations should be made for the sole purpose of gaining general information about fog. If the proposals made below for the study of specific points are carried out, it should be possible gradually to fill in the detail of the general picture.

Under the second heading, quantitative study of the results showed that they could give order-of-magnitude estimates of the physical quantities involved but were not accurate or detailed enough to test or suggest hypotheses about the relations between these quantities. General proposals for more elaborate experimental study of certain points were made in section 3.4 and they are set out more precisely below. These experiments would investigate individual processes which are thought to be important, and should give interesting results, but the experimental study of the conditions leading to the formation of fog could be planned much more effectively if it was guided by a general quantitative theory - however tentative - relating the various factors involved and suggesting what combinations of parameters should be observed. The work of Groen (1947) on night cooling might provide a good starting point for the development of this theoretical background; this work deals only with ground surface conditions and would have to be considerably extended if it was to include changes in the temperature and humidity of the air and so be relevant to fog problems.

Under the third heading, various empirical methods for fog forecasting were examined, using the Cardington data. None of the methods seemed very satisfactory, but it was clear that the data covered too few occasions to afford a reliable test and certainly too few to permit the incorporation of additional parameters into forecasting formulae in the hope of making them more accurate. The data required for this type of investigation are quite different from those required for the study of physical processes; there is no need for detailed data, since it is impracticable to include more than a very few parameters in a statistical treatment, but regularity and uniformity

in the method of observing are prime requirements, and in order to obtain sufficient data in a reasonable time, it is desirable to have observations made at several different stations. The results of the small-scale tests of existing formulae suggest, however, that it might be well to make sure that the best use has been made of existing data (including radio-sonde data) before embarking on a further programme of observation.

5.2. Future programme

Various suggestions for future work have been made in the course of this report. It is obviously beyond the present resources of the Cardington station to carry them all out, and some selection will have to be made. Many of the items have the nature of fundamental rather than applied research and should presumably be considered in relation to the general programme of research in micrometeorology more than in relation to any programme for improving fog forecasts. Even in considering the items directly connected with fog, it must be remembered that they may not contribute very much to the accuracy of forecasting; the major sources of error in fog forecasts are in the forecasting of wind and cloud (see M.R.P. 664, 674 and 718) and will not be affected by the type of work under discussion.

The suggested lines of research are as follows:-

5.2.1. Theoretical work

Attempts should be made to set up equations describing the various processes at work on a radiation night and to produce approximate solutions in typical cases, the aim being to show the effects of interactions between different processes and to suggest important combinations of parameters to be observed. A subsection of this work would be the study of conditions controlling drop growth while a fog is forming, taking account of radiation effects, if necessary.

5.2.2. General study of behaviour of wind, temperature and humidity

5.2.2.1. Peculiarities of wind profiles. The tendency for wind speed to be a maximum at some fairly low height might be worth investigating further. This could best be done through wind and temperature observations at one or two other stations besides Cardington, which should show (a) whether the phenomenon is a local peculiarity and (b) whether it is connected with horizontal temperature gradients.

5.2.2.2. Erratic variations of humidity. It is desirable to make quite certain that the tendency for dew point to vary much more irregularly than temperature (at heights of, say, 300-3000 ft.) is not due to instrumental defects and, if it is not, to decide whether the variations are due to advection. This could be done by a careful study of the instruments and method of using them, at Cardington, and by making "Balthum" ascents at a small network of stations 20 - 50 miles apart.

5.2.2.3. Representativeness of data from one station. In any consideration of the use of data from a single station (particularly for "one dimensional" treatments of physical problems) it is important to know how far the data are typical of wide areas. Direct evidence on this, such as could be obtained from a set of stations at various distances apart, would be valuable.

5.2.3. Study of transfer of heat and water vapour (and momentum) near the ground

This is the central problem in any physical treatment of fog formation. It is obviously closely linked to existing studies (Kew, Cambridge) and is probably best regarded as an extension of them to very stable conditions and to greater heights. There is evidence that extension to heights much above

/100 ft.

100 ft. will be difficult because of advective effects (which may turn out to be serious at smaller heights too) and it is suggested that work should be confined to the lower layers, to start with. In this work it is more important to have complete and accurate data for a few periods of, say, half an hour, than to have data throughout the night on a large number of occasions, which has hitherto been the object of the Cardington programme. The processes needing study are radiation and convection - it can probably be assumed that heat conduction in the ground is well enough understood.

5.2.3.1. Radiation

There is a certain amount of experimental data and fairly convincing theoretical reasons to suggest that estimates of radiation from the usual charts may be seriously in error in typical fog-producing conditions, particularly where the direct cooling of the air by radiation is concerned. This could be investigated by making measurements of (a) total downward radiation at the ground surface and (b) change in net radiative flux between two or more levels.

5.2.3.2. Convection

The aim, in studying convection, is to find relations between the eddy transfer coefficients, K_H and K_W, and the wind and temperature structure and, in particular, to establish how far K_H is equal to K_W. This could probably be achieved by the method attempted in section 3.2 if more accurate data were available. The data required are (a) radiation measurements (as in 5.2.3.1), (b) measurements of heat flow from the earth, (c) measurements of dewfall, (d) measurements of wind, temperature and humidity profiles and rates of change. These data should be obtainable up to heights of 60 ft. with just sufficient accuracy from apparatus already at Cardington (except for (a)). Data could probably also be obtained at one higher level, by using the ballon borne apparatus at a fixed height (150 ft.?) but results obtained, as hitherto, by "spot" measurements at a series of heights are almost certainly not accurate enough for the type of work suggested.

The investigation would be greatly strengthened if, in addition to measurements (a) - (d), measurements could be made of wind drag on the surface and of detailed wind, temperature and humidity profiles in the lowest few feet. The apparatus at present used for such measurements (Cambridge) might need modification for use in nearly calm conditions.

5.2.4. Investigations of fog structure and mode of formation

There are several important problems connected with the properties and behaviour of a fog once it has formed, and study of these problems leads backwards to questions about the details of the way in which fogs form once the relative humidity has reached values near to 100 per cent. The experimental study of these problems requires the measurement of the sizes and other properties of fog droplets, for which no standard apparatus or techniques are available. The first step in the investigation of these problems must therefore be the development of apparatus.

5.2.4.1. Measurement of size, number and composition of fog droplets

Recent interest in the effects of fog, particularly town fog, on health has shown that a knowledge of the size distribution of droplets in typical fogs and, if possible, of their chemical composition, is important in deciding what steps can be taken to safeguard health. This knowledge does not exist, but it should be fairly simple to measure size distribution, once apparatus is developed and tested. These measurements seem likely to lead on to many other investigations into differences between town and country fog, the importance of condensation nuclei, factors affecting stability of fogs and so on, which cannot be clearly foreseen at present.

/5.2.4.2.

5.2.4.2. Variation of visibility with time and height in a fog, and its relation to drop size and number

There appears to be a tendency for visibility in a fog to fall rapidly as the fog forms, then to remain relatively steady, despite considerable fluctuations in temperature. Simultaneous measurements of drop size distribution, liquid water content and visibility should throw light on these changes and perhaps lead towards methods of predicting the value to which visibility will fall.

Variation of visibility with height is discussed in Appendix VI and a tentative explanation of the observations is put forward. This explanation could be tested by measurements of drop size distribution at various heights. If the explanation is correct, it should be possible to determine the height-variation of visibility from ground observations alone, which would be valuable in aircraft operations.

5.2.4.3. Final stages of fog formation

As pointed out in section 3.4 the detailed conditions required for fog formation are not known. Two factors that might be important are radiative cooling of droplets (nuclei) and the mixing of different parcels of air. The matter could be investigated by measuring (a) local variations of temperature, humidity and wind, (b) drop size and number just after fog formation, (c) drop temperature (by radiation measurements or by using fine thermocouples).

5.2.5. Empirical forecasting methods

5.2.5.1. Use of existing data

The rough tests of existing forecasting methods described in section 4.1 suggest that these methods do not make the best use of existing (radio-sonde) data. Adequate statistical tests of the methods should reveal the essential features of each and perhaps show how they could be combined to give an improved method.

5.2.5.2. Collection and use of new data

Although the Cardington results have contributed little to the development of forecasting formulae, it still seems likely that an adequate quantity of data on conditions in the lowest few thousand feet would enable additional parameters to be included in the formulae. These data will only accumulate slowly from a single station particularly if, like Cardington, it is preoccupied with other work, but could be obtained relatively quickly (in a few years) from a network of 4 - 6 stations making simple observations once daily (1800 hr.?).

APPENDIX I

by

R. H. Pedlow (Meteorological Office)

Site and Instruments

1. Site

The site, at 52° 7' N., 00° 25' W., is on level ground in the valley of the Great Ouse at an altitude of 85 ft. above sea level. To the south-east lies a low ridge running NE. - SW., the land rising a little over a mile from the site to nearly 300 ft. at a distance of 2 miles. From south-west to north-east through north, it is at least 5 miles before the land rises above 150 ft., but between NNW. and NNE. only 300 yd. away are two airship sheds 180 ft. high which considerably affect any flow of air from that quarter.

The semi-industrial town of Bedford lies three miles to the north-east, the nearest housing estate being only 1 mile from the site. In open country to the west and south-west there is a large number of brick kilns between 2 and 6 mi. distant.

The immediate surroundings of the site are mostly pasture land on clay soil. A stream runs from west to east in a drainage ditch 6 ft. deep by the north side of the enclosure. In addition to standard instruments with normal exposure, certain other instruments were mounted on two towers and on the cable of a kite-balloon.

2. Instruments

2.1. Standard Met. Office instruments

The following standard instruments were used:-

Dry-bulb, wet-bulb, maximum and minimum thermometers in a large screen.

Grass minimum thermometer.

2-in., 4-in. and 8-in. bent-stem earth thermometers exposed under close-cut grass, not under a bare earth surface.

5-in. raingauge and tilting-siphon rain recorder.

Cup contact anemometer on 20-ft. mast.

Assmann psychrometer.

Pressure-tube anemograph at effective height 135 ft. above ground but 200 ft. above the rain gauge on a 50-ft. rise 1100 yd. to the north-west.

Gold visibility meter with lamps at 800 and 280 yd.

2.2. Balloon-borne instruments

All balloon-borne instruments were fixed by clamps to the flying cable of a kite-balloon. Various balloons have been used but the most commonly used was an HZ-C balloon of 38,000 cu. ft. capacity, capable of lifting the instruments and cables to over 4,000 ft. in light winds. The instruments were fixed about 25 ft. below the belly of the balloon at which distance the airflow suffers little distortion. A 6-volt accumulator supplied current to lamps and to the aspirating motor of the psychrometer. For normal temperature-humidity-windspeed soundings the currents from the instruments to the ground were carried in a steel cable with insulated copper cores (the auxiliary cable) while the balloon was flown on a plain steel cable. When visibility measurements were also required, the balloon was flown on a similiar four-cored cable which was used for the transmission of anemometer

/contacts

contacts and of the high-voltage pulses used in remotely operating switches for the balloon-borne lamps, while the auxiliary cable was then used for thermo-electric and photo-electric currents.

The height of measurements was recorded as length of flying cable paid out; only in a calm is this the true height. When there is any wind the relation between true height and cable length is affected by many variables so that no simple working rule can be established. The ascent was brought to an end if it appeared that the discrepancy between nominal and true height was becoming excessive. Comparison with simple cases and measurement from sketches of the shape of the cable indicate that when this decision was reached the maximum error was probably about 10 per cent. For measurements within the strata of light winds on radiation nights the error was almost certainly less than 5 per cent.

2.2.1. Balloon-borne psychrometer

The balloon-borne psychrometer differed in two important particulars from that described by Swinbank (1946(1)), viz., it was not a quick-response instrument and consequently was not normally adapted for automatic recording. The thermocouples were made of 22 S.W.G. wire giving a 90 per cent response to sudden temperature change in about 3 minutes. As the e.m.f's generated were transmitted down the auxiliary instead of the flying cable, the copper-mercury contacts of Swinbank's circuit were avoided.

Calibration was carried out at intervals of several months with the junctions in water. The e.m.f. was about 0.08 mV per 1°F, and could be read to an accuracy of ±0.01 mV. Measurements taken on two occasions when the balloon was left stationary at fixed heights of 2000 and 500 ft. in radiation conditions indicated that despite the smoothing effect of the slow-response thermocouples, sampling errors were not negligible. Dry-bulb changes of the order of 1°F in 5 to 10 minutes were noted, and wet-bulb depressions were even more erratic in fairly dry air, one depression, averaging about 3.5°F, changing by 1.7°F in 5 minutes. There was nothing to indicate that this was due to faulty irrigation or aspiration of the wet-bulb and it is probable that sampling errors of 25 per cent in wet-bulb depressions are not uncommon, though further investigation is required.

2.2.2. Balloon-borne anemometer

The balloon-borne anemometer during the 1951-2 and 1952-3 seasons was a cup contact anemometer Mk. I with a stopping speed of about 2 kt. and an accuracy of ±1 kt. throughout its range. In the 1953-4 season this was most frequently replaced by a small sensitive cup anemometer made by Casella with a stopping speed of 1 ft./sec. and accuracy (limited by the reading period) of 0.2 ft./sec. In order to keep the duration of an ascent within reasonable limits windspeeds were read over one minute at each height. It was established that the probable error of a one-minute sample was less than 1 kt. The increase in windspeed due to the proximity of the balloon is of the order of 2 per cent.

2.2.3. Fog-density indicator

The fog-density indicator consists of 24 selenium photo-electric cells, each 2 sq. in. in area, connected in parallel and sandwiched between rectangular Perspex sheets with half the cells facing each way. In a rectangular space in the middle of the instrument three 6v. 12w. lamps are mounted in the same plane as the cells which are protected from the direct light of the lamps by a metal strip. When the lamps are lit in a turbid atmosphere, the light is reflected from suspended particles on to the cells and the resultant photoelectric current is a measure of the scattered light. The indicator can, of course, be used only in darkness. Calibration was carried out against the photo-electric visibility

/meter

meter described in para. 2.3.3 below, with both instruments at 55 ft. to avoid reflection from the ground to the indicator. A relationship:-

$$iV = 284$$

was found for the range 300 - 1200 yd., where i is the photo-electric current in microamps and V is the visibility in yards. As currents of the order of 0.5 microamp are caused in clear air by bright moonlight and at times (apparently) by stray loops of leads, the instrument is best regarded at present as an indicator of the presence of fog with visibility less than about 440 yd., though a cleaning up of the wiring may make possible the doubling of this distance.

2.2.4. Slant visibility

Measurements of slant visibility were made by walking away from the balloon to the extinction point of a 6v. 12w. (13 c.p.) lamp hanging below the other balloon-borne instruments.

2.3. Tower instruments

Variations in the vertical of temperature, humidity, windspeed and visibility were measured using two 55-ft. towers of tubular steel scaffolding built 160 yd. apart in a N.-S. line. The N. tower is about 5 ft. square at the base tapering to 3 ft. square at the top and has wooden platforms within the tower at about 17 and 51 ft. The S. tower has so far been used only as a support for lamps at 4 and 55 ft. (see para. 2.3.3 below).

2.3.1. Tower psychrometers

At 4, 20 and 55 ft. on the N. tower were aspirated wet- and dry-bulb thermocouples, each consisting of copper and constantan wires soldered to a cylindrical copper 'bulb' 1.0 in. long and 0.2 in. diameter, the bulb being fixed to the end of a hollow 3 in. rod of Tufnol supported in a half-inch-square air duct. Each pair of ducts (wet- and dry-bulb) was aspirated through a common orifice by a centrifugal air-pump driven by a 24 v. motor. The circuit was wired so that three such pairs of thermocouples plus a single junction in a vacuum flask of crushed ice and water provided measurements of all required temperatures, the effect of the electrically more remote thermocouples on the various galvanometers having been shown to be negligible. All the galvanometers recorded on the same sheet of bromide paper and time-marks were made automatically every 20 minutes. Calibration was carried out with the thermocouples in water and the galvanometers adjusted so that that which measured the temperature difference between the 4-ft. dry-bulb and ice had a deflexion of 0.5 cm/1°F, while those measuring the 4-ft. to 20-ft. and 20-ft. to 60-ft. dry-bulbs and all the dry- to wet-bulb differences gave deflexions of 1.0 cm/1°F. With this arrangement, temperature differences correct to 0.1°F should be measurable.

Accurate measurement, however, was made impossible by leakage from the aspirating motors and other 12- and 24-v. D.C. circuits on the tower into the thermocouple circuits, introducing errors of the order of 1°F which were aggravated in fog conditions, It has proved impossible to isolate the surface leaks and the trouble is now being obviated by the use of A.C. in all such circuits.

2.3.2. Tower anemometers

Three Casella sensitive anemometers, at 4, 20 and 55 ft., were used, each at the SW. corner of the N. tower, the two lower instruments being set out 3 ft. from the tower. Such small interference as the tower has on airflow thus affected the anemometers only when the wind came from the northerly quarter where it is already affected by the airship sheds. The

/recorder

recorder was essentially three recorders of the type described by Bibby (1944), with the impulse motors mounted one above the other, the three pens recording on one drum which was continuously driven by clockwork instead of being moved at intervals by another impulse motor. The pens were returned to zero at 20-min. intervals. The scale varied according to the anemometer in use between 6 and 8 ft./sec. per inch of trace. The stopping speed of the anemometers was 0.5 ft./sec., and the traces were read to the nearest 0.5 ft./sec., well within the accuracy of the instruments.

2.3.3. Photo-electric visibility meter

Photo-electric visibility meters were in use at 4 and 55 ft. The beam from a 24v. 250w. signal lamp in a parabolic reflector was directed over a 160-yd. range and the lamp image focussed by a 5 in. diameter lens of focal length 36 in. on to a selenium cell. Baffles within the tube of the instrument restricted the stray light which could reach the cell, and the field of view was so small that the meter could be used in full daylight except that the scatter of bright sunshine on dispersing fog caused fluctuations of photo-electric current so rapid that no useful measurements could then be made of the photographic trace from the galvanometers. The lamps were switched off for 40 seconds every 15 minutes, giving a measure of the deflexion due to daylight. A heater inside the meter lens prevented condensation from forming on the outer surface.

Since the brightness of a lamp varies with the third power of the applied voltage, the mains supply was transformed down, rectified and passed through a shunted galvanometer so that it could be monitored on the same bromide as the record from the meters, and the photo-electric current corrected accordingly.

Where

i_o = galvo. deflexion for standard voltage and infinite visibility,
i = " " " " " " visibility V,
and d = distance between lamp and cell,

$$V \log_e (i_o/i) = 3.91 d.$$

The galvanometer was so arranged that i_o was approximately 8 cm. The uncertainty in the measurement of the mains-monitor trace of about 1 per cent and in the measurement of the photoelectric deflexions of about 1 mm. combined to give an uncertainty of about ± 10 per cent at 1100 yd. visibility, ± 7 per cent between 900 and 400 yd., ± 5 per cent at 300 yd. and ± 11 per cent at 225 yd. During observations, whenever other work permitted, the 160-yd. lamp was switched off and replaced by a 230v. 300w. lamp at 30 yd. from the meter when the visibility fell below 250 yd., the galvanometer being simultaneously changed for one which maintained i_o at about 8 cm. (This arrangement was only possible with the 4-ft. meter). The accuracy was still no better than ± 10 per cent above 170 yd., but was about ± 5 per cent for lower visibilities, down to about 30 yd.

Comparison of two cells working from the same lamp showed that small variations in the contacts at the cells introduced an uncertainty of about ±2 per cent into the value of i_o, which implies a further uncertainty of about ± 3 per cent in a visibility of 1100 yd., decreasing to ± 1 per cent at 250 yd. using the 160-yd. lamp, and about ± 3 or 4 per cent below 250 yd. using the 30-yd. lamp. Thus the total uncertainty was of the order of ± 150 yd. at 1100 yd. decreasing to ± 15 yd. at 250 yd.; or, using the 30-yd. lamp, ± 35 yd. at 250 yd. and ± 8 yd. at 100 yd. Considerably larger departures from eye observations occurred when the visibility was patchy and/or steeply stratified, and occasional gross errors arose when shifting contacts at either the cell or the lamp altered the value of i_o.

/2.3.4.

2.3.4. Wind direction recorder

A distant-reading wind-direction recorder was exposed at 56 ft. on the N. tower.

2.4. Earth thermocouples

Occasional readings were made with thermocouple earth thermometers with junctions at $\frac{1}{2}$, 1, 2, 4, 8 and 16 in. below ground level, visual measurements of temperature difference from a junction in ice being made by potentiometer at hourly intervals. Temperatures correct to 0.1°F could be read. The electrical insulation of the thermocouples from the ground presented difficulty and the failure of several types of insulation interrupted the readings.

2.5. Dew measurements

After preliminary experiment with a recording dew-balance and Duvdevani dew gauges, both of which were discarded as unreliable, dew measurements were made using the technique employed by Pasquill (1949) except that the evaporimeter pans and outer containers were, during the 1953-4 season, fitted with bottoms of thin sheet brass instead of Tufnol, so that the heat exchange with the ground beneath should more nearly resemble that in the surrounding soil. Originally six, but in the last season four, pans were weighed, at first irregularly but ultimately at hourly intervals, differences being measured to ± 0.05 gm.

There was the same sort of scatter between the differences shown by the various evaporimeter pans as experienced by Pasquill (loc. cit.) and Rider (1954). The standard error of the mean difference was found for fifty separate hourly periods covering four different occasions. The average value of these standard errors was equivalent to 0.010mm. dewfall, but this was aggravated by a few exceedingly high values (the highest 0.085 mm.). The median value was 0.006 mm., the interquartile range being 0.004 to 0.011 mm.

APPENDIX II

Accuracy required in estimates of radiation

If we try to predict the phenomena of cooling and fog formation on a clear night by calculation of the physical processes involved, radiation enters the calculation in two ways. In the first place it is an important factor determining the temperature of the ground surface - the boundary condition for heat flow in earth and air. In the second place, we have seen that direct radiative cooling of the air is at least as important as convective cooling in determining the air temperature.

The surface temperature is fixed by the condition that the net heat flux must be zero; if we assume that the conductive and convective fluxes and the downward flow of radiation are fixed, the surface temperature will adjust itself until the outward radiation (equal to the radiation from a black body at the temperature of the surface) has the value required to balance the heat budget. This outward radiation changes by about 0.5 mW/cm^2 for each degree centigrade, in the range of temperature we are concerned with, so that if we wish to determine the surface temperature to 1°C, we must know each flux component, including the downward flux of radiation from the atmosphere, to 0.5 mW/cm^2. We have seen (Table II) that the downward radiation is usually of the order of 25 mW/cm^2, so that it is required to estimate it to about 2 per cent (assuming that the standard of accuracy implied by 1°C error in surface temperature is reasonable - some evidence for this is provided by fig. 16, section 4.2).

Robinson (1947) has shown that results of the required order of accuracy can be obtained from radiation charts on many occasions, though the presence of "additional radiation" may sometimes cause larger errors, and his simple treatment of carbon dioxide may be inaccurate in conditions with large temperature gradients. It is not difficult, using the Kew chart, to replace the simple assumption that infinitely thin layers of CO_2 are "black" within their absorption band by a more realistic estimate of the variation of emissivity with path length, provided it is assumed that the absorption regions of CO_2 and water vapour do not overlap. This was done for the estimates of radiation on 27.9.53 (see Table II); downward radiation was found to be 0.3 mW/cm^2 greater with the more exact treatment of CO_2 than with the simple treatment. The effects of radiation from suspended particles may well be more serious, but are difficult to estimate exactly; a rough treatment can be given, as follows.

To estimate downward radiation, using the Kew chart, (see fig. 17) we plot the percentage (of black body) emissivity of the layer of atmosphere up to a given height against the black body radiation corresponding to the temperature at that height. The emissivity is assumed to be a function of the total amount of water in the layer (modified by a pressure correction) only. If the atmosphere contains smoke or mist particles, assumed to radiate as black bodies,* the emissivity will be modified, If the particles in a given layer are sufficient to obscure a fraction A of the sky, we may suppose that the emissivity of the layer is made up of two parts, a fraction A

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* It is by no means certain that all aerosol particles are "black", or that a very small particle, even of black material, will radiate as a black body at wavelengths large compared with its own size. In the absence of any full theoretical discussion of these points or of relevant experimental evidence, this Appendix makes the simplest assumption, that all the particles radiate as perfect black bodies. The estimates obtained for the effects of radiation from particles thus represent an upper limit.

representing the black body radiation from the particles and a fraction $(1 - A)\ (E(u) + E(CO_2))$ representing radiation from water vapour in the parts not screened by particles ($E(u)$ being the emissivity of the water vapour, with path length in the layer considered and $E(CO_2)$ the emissivity of the carbon dioxide). The total emissivity of the layer is therefore $E(u) + E(CO_2) + A(1 - E(u) - E(CO_2))$. On the Kew chart $E(u) + (E(CO_2)$ is represented by the distance between the base-line and the plotted curve (assuming $E(CO_2)$ to have a constant value of 0.185, whatever the thickness of the layer in this rough treatment) and $(1 - E(u) - E(CO_2))$ by the distance from the curve to the top of the diagram. The modification produced by particle radiation is therefore to move each point on the curve towards the top of the diagram by the fraction A of its original distance from the top, the result being shown by the uppermost curve of fig. 17.

The fraction A can be estimated from the visibility. If the attenuation coefficient for visible light is β, then the "visibility" is approximately $4/\beta$ (Koschmieder's formula) and the fraction A is $1-\exp(-\beta z)$ for a layer of height z, assuming that β is independent of height. On 29.9.51 (fig. 1) the visibility at ground level was about 2500 yd. If we assume that the visibility was constant up to 500 ft. (the height of the inversion) and that the air above was clear, we get values of A increasing from 0.005 at 100 ft. to 0.22 at 500 ft. and remaining constant for greater heights. These values were used to produce the modified curve of fig. 17 and so to calculate the extra downward radiation to be expected from the smoke or mist particles that were restricting the visibility. The area between the two curves of fig. 17 corresponds to extra downward radiation of about 2 mW/cm^2 - a very significant amount. On an occasion such as 5.12.52, when visibility was less and water vapour concentration smaller, the particle radiation must have been still more important. If particle radiation is as important as these rough estimates suggest, it should be measurable by ordinary radiation instruments, in suitable conditions, and it seems desirable to make such measurements before trying to develop the theoretical treatment further.

Turning to the calculation of the rate of cooling of the air by direct radiation, it is clear from the figures of Table II that the radiative cooling term is very sensitive to errors in the calculation of radiative fluxes. The cooling is, in fact, measured by long thin areas on a radiation chart (as explained by Robinson, 1950) which are sensitive to errors in the chart co-ordinates (or in the actual plotting of points) and are difficult to measure accurately. Some of these difficulties can be avoided by using a "flux divergence chart" in which, effectively, the heights of the long thin areas on the ordinary radiation chart are plotted against their horizontal co-ordinate, to give better shaped areas to measure. However, the co-ordinates of such a chart (in the simplest case dE/du and σT^4, E being fractional emissivity of a layer with water path u and σT^4 the black body radiation at temperature T) depend on the rate of change of emissivity with thickness of air-layer instead of on the emissivity itself, as in the more usual charts, and since this rate of change is known much less accurately than the emissivity, a good deal of the apparent gain in precision is illusory. Nevertheless, the plotting of data on a flux divergence chart (fig. 18) does help to show more clearly some of the factors controlling radiative cooling. Fig.18 represents conditions at 100 ft. height at 0100 hr. on 27.9.53. The area to the left of the central vertical line represents the net exchange of the layer at 100 ft. with the ground and the air up to 100 ft.; the area to the right represents the exchange with layers above 100 ft. The temperature gradient is positive upwards at 100 ft. so the left hand area of fig. 18 represents cooling, the right hand area heating, and the difference of the two areas represents the net cooling effect. Near the central line (representing the 100 ft. level in this case) the curves rise to very high values, corresponding to the large value of dE/du for short water paths; the curves are not well determined in this region because the temperature observations were not close enough and because the values of dE/du are uncertain. If there were any marked discontinuities in temperature gradient near 100 ft., areas near the central line in fig. 18 might contribute very large amounts to the area

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representing net heating or cooling, but if we assume that short-range radiative exchange is such an efficient transfer mechanism that it smooths out such discontinuities, it is reasonable to assume that areas to the right and left of the central line cancel each other out in the immediate neighbourhood of the line, leaving only the areas below the (arbitrarily drawn) broken horizontal line to be taken into account.

Areas on fig. 18 actually represent the radiative heating or cooling of a layer containing unit mass (per unit area) of water vapour, but this can, of course, be expressed as the rate of change of temperature by radiation, knowing the water vapour content and specific heat of the air at the height in question. The result, from fig. 18, is that radiative exchange with the layers below 100 ft. produces a cooling of 2.4×10^{-4} °F/sec. and exchange with the layers above a heating of 0.8×10^{-4} °F/sec. giving a net cooling rate of 1.6×10^{-4} °F/sec. Fig. 18 suggests that a rough approximation to the radiative cooling at levels not far below the level of maximum temperature, in inversion conditions, could be obtained by multiplying the difference between air temperature and surface temperature by a function of the water path from the surface to the level in question; it is clear that the surface temperature is one of the most important factors determining the cooling rate.

Although the emissivity data for water vapour are inadequate to determine the vertical co-ordinate scale of charts such as fig. 18 at all precisely, the biggest source of error in radiative cooling estimates is probably the presence of radiation from suspended particles. Modification of fig. 18 to take account of this is possible but rather complicated and a simpler procedure will serve to show the importance of the effect.

We consider again the conditions at 0001 hr. on 29.9.51 (figs. 1 and 17), supposing that a hazy layer with visibility 2500 yd. extended up to 500 ft., and will calculate the cooling due to radiation from the particles in the layer from the ground to 100 ft. (which can be compared with the figures in Table II - particularly the cooling of 0.3 mW/cm^2 by radiation from water vapour and carbon dioxide). If visibility was 2500 yd. the attenuation coefficient (the projected area of scattering particles per unit volume) must have been of the order of 4/2500 yd. or 1/2000 ft. The particles in a 1 cm^2 column of the lowest 100 ft. therefore had a projected area of 1/20 cm^2 and emitted 1/20 of the black body radiation (per cm^2) appropriate to their temperature in each direction (1/20 upwards, 1/20 downwards). However, they also absorbed 1/20 of the upward flux of radiation and 1/20 of the downward flux. The mean temperature between 0 and 100 ft. was 8.5°C, the black body radiation appropriate to this temperature is 35.5 mW/cm^2, the upward flux was 34.0 mW/cm^2 and the downward flux 27.2 mW/cm^2 (from Table II). These fluxes were calculated neglecting the presence of the particles; we have already shown that the particles would increase the downward flux at the surface by about 2 mW/cm^2, and for this rough calculation we shall assume that this applies to the whole of the lowest 100 ft. and is the only correction that need be made. The net radiation from the particles in a 1 cm^2 column 100 ft. high is, therefore,

$$\frac{1}{20}\left(2 \times 35.5 - 34.0 - 27.2 - 2\right) \text{mW} \approx 0.4 \text{ mW}$$

which is actually greater than the net radiation from water vapour and carbon dioxide in the same layer. (We may note that a cooling of 0.4 mW/cm^2 in a 100 ft. layer corresponds to a rate of fall of temperature of about 0.6 °F/hr).

APPENDIX III

Heat flow from freezing ground

The purpose of this appendix is not to give exact solutions (for which reference may be made to Ingersoll, Zobell and Ingersoll, 1948) but to estimate orders of magnitude in typical cases. We investigate the quantity of heat flowing to the surface when the surface temperature is held at a fixed amount (ΔT) below the initial earth temperature (supposed uniform with depth) in two cases, first with the surface temperature above freezing point, secondly with the surface temperature below freezing point. In the first case the well-known solution for the heat flow, H, at time t is

$$H = \Delta T \left(\frac{K \rho C}{\pi t} \right)^{\frac{1}{2}}$$

where K, ρ , C are the thermal conductivity, density and specific heat, respectively for the soil. If we use the typical values ΔT = 5°C, K = 0.002 cal.cm^{-1}sec.$^{-1}$ °C^{-1}; ρ = 1.5 gm.cm^{-3}. C = 0.5 cal.°C^{-1}gm.$^{-1}$

we find $H = 7 \times 10^{-4}$ cal.cm^{-2}sec.$^{-1}$

$\equiv 2.9$ mW/cm^2

for t = 3 hr., agreeing roughly with the heat fluxes found in Tables I and II.

In the second case we simplify matters by assuming the earth to be just above freezing point initially. Then a sudden fall in temperature, ΔT , at the surface will freeze the earth from the surface downwards. We suppose that at time t freezing has penetrated to a depth f. Then the temperature gradient in the upper layers will be approximately $\Delta T/f$ and the heat flow $K.\Delta T/f$ (neglecting the heat lost by these layers in comparison with the heat flowing through them). This heat flow can be equated to the rate of release of latent heat as the freezing level falls;

$$H = K.\Delta T/f = w L \, df/dt$$

where w is the water content of the soil (gm./cm^3) and L the latent heat of fusion of ice. We thus obtain

$$f = \left(\frac{2 K.\Delta T.\, t}{w L} \right)^{\frac{1}{2}}$$

$$H = \left(\frac{K w L.\Delta T}{2t} \right)^{\frac{1}{2}}$$

Taking K = 0.002, ΔT = 5, t = 3 hr. as before, with w = 0.5 gm./cm^3 and L = 80 cal./gm. we find

$f = 2.2$ cm

$\frac{df}{dt} = 1.1 \times 10^{-4}$ cm /sec. (approximately 1 cm./3 hr.)

$H = 45.5 \times 10^{-4}$ cal.cm $^{-2}$ sec.$^{-1}$

$\equiv 19$ mW/cm^2

In our hypothetical conditions, therefore, the heat flow is some six times greater in the freezing than in the non-freezing case, for a given fall in

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surface temperature. In practice, of course, the greater heat flow serves to increase the surface temperature, so that a state of balance is reached on freezing nights, other things being equal, with a smaller earth-air temperature difference than on non-freezing nights.

We may note that, in the conditions considered, the rate of loss of heat from the earth above the freezing level (2.2 cm. at t = 3 hr.) is about 0.8 mW/cm^2 so that its neglect in comparison with the flow of 19 mW/cm^2 was justified for a first approximation. If the earth is initially above freezing point, the above treatment will still be roughly correct, if ΔT is taken to be the amount by which the surface temperature falls below freezing point; there will be an extra contribution to the heat reaching the surface, from the layers below the freezing level, but this will be comparable to the heat flow in the "non-freezing" case and will be small compared with the latent heat contribution unless ΔT is very small.

APPENDIX IV

Effect of convection on relative humidity

The equation for the rate of change of temperature at any level by convection is

$$\frac{d\theta}{dt} = K_H \frac{\partial^2\theta}{\partial h^2} + \frac{\partial K_H}{\partial h}\frac{\partial \theta}{\partial h} \qquad (1)$$

where θ is potential temperature, t time, h height and K_H the eddy diffusion coefficient for heat. A similar equation governs the rate of change of specific humidity

$$\frac{dw}{dt} = K_w \frac{\partial^2 w}{\partial h^2} + \frac{\partial K_w}{\partial h}\frac{\partial w}{\partial h} \qquad (2)$$

where w is the specific humidity (mass of water vapour per unit mass of air) and K_W the eddy diffusion coefficient for water vapour. On a radiation night, convection leads to a fall in temperature and a fall in water content near the ground. There seems little prospect of predicting these falls with any accuracy from (1) and (2) because in most cases the terms on the right of the equations are nearly equal and of opposite sign, so that very accurate knowledge of the K's and their variation with height would be required to find the left-hand terms precisely. Thus, in a typical case from Table II (27.9.53) we have, at 100 ft.,

$$\frac{d\theta}{dt} = -1\times10^{-4} \ ^\circ\text{C sec}^{-1}$$

$$K_H \frac{\partial^2\theta}{\partial h^2} = -8\times10^{-4} \ ^\circ\text{C sec}^{-1}$$

$$\frac{\partial K_H}{\partial h}\frac{\partial\theta}{\partial h} = 7\times10^{-4} \ ^\circ\text{C sec}^{-1}$$

However, in the problem of fog formation we are not primarily interested in the temperature or the water content separately; the quantity of interest is the nearness to saturation, conveniently expressed by the difference between the saturation and actual specific humidities and written $\Delta\ (=w_s - w)$. It is reasonable to hope that $d\Delta/dt$ may be expressible in a form which depends less critically on the exact values of the K's, provided there is some simple relation between K_H and K_W. We assume the simplest relation $K_H = K_W$ (= K, say) and try to find an expression for $d\Delta/dt$. We have

$$\frac{d\Delta}{dt} = \frac{dw_s}{dt} - \frac{dw}{dt}$$

$$= \left(\frac{\partial w_s}{\partial\theta}\right)_h \frac{\partial\theta}{\partial t} - \frac{dw}{dt}$$

$$= \left(\frac{\partial w_s}{\partial\theta}\right)_h \left(K\frac{\partial^2\theta}{\partial h^2} + \frac{\partial K}{\partial h}\frac{\partial\theta}{\partial h}\right) - \left(K\frac{\partial^2 w}{\partial h^2} + \frac{\partial K}{\partial h}\frac{\partial w}{\partial h}\right)$$

which can be rearranged by using the relations

$$\frac{\partial w_s}{\partial h} = \left(\frac{\partial w_s}{\partial h}\right)_\theta + \left(\frac{\partial w_s}{\partial \theta}\right)_h \frac{\partial \theta}{\partial h}$$

$$\frac{\partial^2 w_s}{\partial h^2} = \left(\frac{\partial^2 w_s}{\partial h^2}\right)_\theta + \left(\frac{\partial^2 w_s}{\partial \theta^2}\right)_h \left(\frac{\partial \theta}{\partial h}\right)^2 + \left(\frac{\partial w_s}{\partial \theta}\right)_h \frac{\partial^2 \theta}{\partial h^2} + 2\frac{\partial^2 w_s}{\partial \theta \partial h} \cdot \frac{\partial \theta}{\partial h}$$

to give

$$\frac{d\Delta}{dt} = K\frac{\partial^2 \Delta}{\partial h^2} + \frac{\partial K}{\partial h}\frac{\partial \Delta}{\partial h} - K\left(\frac{\partial^2 w_s}{\partial h^2}\right)_\theta - \frac{\partial K}{\partial h}\left(\frac{\partial w_s}{\partial h}\right)_\theta$$

$$- K\frac{\partial \theta}{\partial h}\left[\left(\frac{\partial^2 w_s}{\partial \theta^2}\right)_h \frac{\partial \theta}{\partial h} + 2\frac{\partial^2 w_s}{\partial \theta \partial h}\right] \qquad (3)$$

which expresses the rate of change of Δ by an ordinary diffusion equation, modified by correcting terms for the variation of saturation specific humidity with height (at constant temperature) and for the curvature of the relation between saturation specific humidity and temperature. If it turns out that the individual terms of (3) are not large compared with their algebraic sum, then (3) will give a more accurate value of Δ than would be obtained by using (1) and (2) separately.

In particular we may be able, without introducing too much error, to eliminate $\partial K/\partial h$ from (3) by using the approximate relation

$$\frac{\partial \theta}{\partial h}\frac{\partial K}{\partial h} = -K\frac{\partial^2 \theta}{\partial h^2} \qquad (4)$$

Using this relation, (3) becomes

$$\frac{d\Delta}{dt} = K\left[\frac{\partial^2 \Delta}{\partial h^2} - \frac{\partial^2 \theta/\partial h^2}{\partial \theta/\partial h}\frac{\partial \Delta}{\partial h}\right]$$

$$- K\left[\left(\frac{\partial^2 w_s}{\partial h^2}\right)_\theta - \frac{\partial^2 \theta/\partial h^2}{\partial \theta/\partial h}\left(\frac{\partial w_s}{\partial h}\right)_\theta\right]$$

$$- \frac{\partial \theta}{\partial h} K\left[\left(\frac{\partial^2 w_s}{\partial \theta^2}\right)_h \frac{\partial \theta}{\partial h} + 2\frac{\partial^2 w_s}{\partial \theta . \partial h}\right] \qquad (5)$$

Equation (5) can be used to find $d\Delta/dt$ provided we know K and have sufficiently accurate profiles of temperature and humidity. Although

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the Cardington data are far from accurate enough it is interesting to use them in (5) to illustrate the magnitude of the various terms.

From fig. 2 and Table II we obtain, for 0100 hr. on 27.9.53 at 100 ft.,

$$K = 3800 \text{ cm}^2 \text{ sec}^{-1}, \quad \partial\theta/\partial h = 5.2 \times 10^{-4} \text{ °C cm}^{-1},$$
$$\partial^2\theta/\partial h^2 = -1.9 \times 10^{-7} \text{ °C. cm}^{-2}, \quad \partial\Delta/\partial h = 9.1 \times 10^{-8} \text{ cm}^{-1},$$
$$\partial^2\Delta/\partial h^2 = -9.0 \times 10^{-12} \text{ cm}^{-2}$$

(The variation of Δ with h given by the balloon ascents is very irregular, and drastically smoothed curves have been used to obtain the above values). The values of the constants in eq. (5) are as follows (at 10°C):

$$\left(\frac{\partial^2 w_s}{\partial h^2}\right)_\theta = +2.1 \times 10^{-13} \text{ cm}^{-2}; \quad \left(\frac{\partial w_s}{\partial h}\right)_\theta = -3.8 \times 10^{-8} \text{ cm}^{-1};$$

$$\left(\frac{\partial^2 w_s}{\partial \theta^2}\right)_h = +3.2 \times 10^{-5} \text{ °C}^{-2}; \quad \frac{\partial^2 w_s}{\partial\theta.\partial h} = -2.3 \times 10^{-9} \text{ cm}^{-1} \text{ °C}^{-1}.$$

Putting these values in eq. 5 we obtain

$$\frac{10^{12}}{3800}\cdot\frac{d\Delta}{dt} = -9.0 + 33.2 - 0.21 + 13.9 - 8.7 + 2.4$$

(the terms are arranged in the same order as in eq. (5).)

$$= 31.6 \text{ cm}^{-2}$$

$$\frac{d\Delta}{dt} = 1.2 \times 10^{-7} \text{ sec.}^{-1}$$

$$\equiv 1.2 \text{ gm./Kg. in } 10^4 \text{ seconds.}$$

We notice that the second term is numerically a good deal greater than the first and the fourth considerably greater than the third, so that the effects of the approximation made by using (4) will not be too serious. However, the final result does not agree at all well with the observations; the actual value of $d\Delta/dt$ should be the sum of the part due to convection, which we have just calculated, and a part due to radiative cooling (if the air is cooled by radiation the saturation specific humidity, w_s, will fall and Δ will therefore decrease), but the values in the case we are considering were

Convection contribution to $d\Delta/dt$	$= 1.2 \times 10^{-7}$ sec.$^{-1}$
Radiation " " $d\Delta/dt$	$= -0.3 \times 10^{-7}$ sec.$^{-1}$
Observed value of $d\Delta/dt$	$= -0.2 \times 10^{-7}$ sec.$^{-1}$

Equation (5) thus appears to overestimate the effects of convection very considerably, but the possible errors in the quantities we have had to use are so large that they could easily account for the discrepancy, and we must conclude that it is impossible to test the validity of (5) without much more accurate observations. The chief sources of error are in the estimates of radiation, which we have already discussed, and in the profiles of humidity.

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In order to test (5) to an accuracy of the order of 20 per cent we should require radiation measurements or estimates to about 0.1 mW/cm^2 and smooth humidity profiles defined to about 0.2°F in dew point - if the humidity profiles turn out to be irregular even at small heights and when averaged over reasonable periods, it is clear that there will be little possibility of dealing with humidity changes in the way attempted here.

Just before fog forms, the air is often very close to saturation for some distance above the ground, and it is interesting to apply eq. (5) to the ideal case with $\Delta = 0$. If $d\Delta/dt$ is found to be negative it would indicate the onset of supersaturation, and we should expect fog to form; if it is positive the air would be becoming drier (unless radiation counteracted this effect) and no fog should form. From (5) we see that with $\Delta = 0$ the limiting case $d\Delta/dt = 0$ occurs when

$$\frac{\partial^2\theta/\partial h^2}{\partial\theta/\partial h}\left(\frac{\partial w_s}{\partial h}\right)_\theta = \left(\frac{\partial^2 w_s}{\partial h^2}\right)_\theta + \left(\frac{\partial^2 w_s}{\partial\theta^2}\right)_h \left(\frac{\partial\theta}{\partial h}\right)^2 + 2\,\frac{\partial^2 w_s}{\partial\theta\,\partial h}\,\frac{\partial\theta}{\partial h} \qquad (6)$$

independently of K. Putting in the values of constants used before, (6) becomes

$$-3{\cdot}8\times10^{-8}\,\frac{\partial^2\theta}{\partial h^2} = 3{\cdot}2\times10^{-5}\left(\frac{\partial\theta}{\partial h}\right)^3 - 4{\cdot}6\times10^{-9}\left(\frac{\partial\theta}{\partial h}\right)^2 + 2{\cdot}1\times10^{-13}\left(\frac{\partial\theta}{\partial h}\right) \qquad (7)$$

If, as an illustration, we put $\partial^2\theta/\partial h^2 = -1{\cdot}9\times10^{-7}$ °C.cm^{-2} (as at 0100 on 27.9.53) we can solve (7) approximately for $\partial\theta/\partial h$ and find

$$\partial\theta/\partial h \simeq 0{\cdot}7\times10^{-3}\ °\text{C.cm}^{-1}$$

This implies that, with the postulated value of $\partial^2\theta/\partial h^2$ and at a temperature of about 50°F, the relative humidity of an already practically saturated atmosphere should increase if there is an inversion steeper than 0.7 x 10^{-3} °C/cm., or about 4°F/100 ft., and decrease if the inversion is less steep, unless radiation or other effects outweigh the convection process.

APPENDIX V

Effects of radiation on growth of fog droplets

It is well known (eg. Best, 1954) that the size of a droplet of salt solution varies in a characteristic way with the relative humidity of the surrounding air. As the humidity is increased the equilibrium size of droplet increases, slowly at first then more rapidly until, for some value of relative humidity in excess of 100 per cent, the size will increase without limit, if the humidity is held constant. It is tempting to identify this phase of unlimited growth with the rather sudden transition which we have seen often occurs from mists with visibilities of the order of a mile to fogs with visibility of the order of 100 yd. It must be borne in mind, however, that the theoretical unlimited growth can only occur if the relative humidity of the atmosphere remains constant, and that the growth of a cloud of drops will tend to reduce the relative humidity both by removing water vapour and by raising the temperature, through the release of latent heat. We find, in fact, that if we replace the condition of a fixed relative humidity by the more realistic one of a fixed temperature and take reasonable values for the size and concentration of droplet nuclei, there is no stage where unlimited growth is possible. As the temperature is lowered the drops will grow, but the total amount of water condensed is always fixed by the temperature, being, to a close approximation, equal to the difference between the original vapour density ($gm./m^3$) and the saturation vapour density at the temperature reached.

It seems possible to account for the sudden growth of droplets by allowing for the effects of radiation. As we saw in Appendix II, a particle 100 ft. above the ground on a typical radiation night would lose heat by radiation at a rate of the order of 8 mW/cm^2 of its area projected on a horizontal plane. As a consequence the particle will be at a slightly lower temperature than its surroundings, so that the loss of heat by radiation is compensated by heat conducted from the surrounding air. In the case of a droplet, its lower temperature will mean that it will be in equilibrium with air of a slightly lower humidity than a droplet of the same size not subject to radiative cooling. This effect is quite small, as is shown by Table V, calculated for a droplet formed on a sodium chloride nucleus of mass 6×10^{-14} gm.

Table V

Equilibrium radius of droplet (microns)	1.0	1.2	1.5	1.9	2.5	3.1
Relative humidity, without cooling per cent	99.99	100.04	100.04	100.05	100.04	100.04
Relative humidity, with cooling at 8 mW/cm^2 per cent	99.98	100.02	100.03	100.03	100.01	100.01

Table V suggests that radiative cooling will have little direct effect on the early stages of drop growth. When the critical size (corresponding to the underlined values of R.H. in Table V) is reached, radiation effects become important. Condensation will occur freely so long as the relative humidity remains at or above the values in the table (i.e. for most practical purposes at 100 per cent), but, as already pointed out, condensation tends to lower the relative humidity. When the drops are radiating, the rate of condensation can be found by equating the loss of heat by radiation to the sum of the latent heat of condensation and the heat that must be removed from the air (via the drop) to keep it saturated in spite of the loss of water vapour. For a drop of radius r we can write the radiative loss of heat as $\pi r^2 R$,

/the

the latent heat of condensation as $4\pi r^2 L\, dr/dt$ and the heat to be extracted from the air in order to maintain saturation as

$$\frac{4\pi r^2 dr/dt}{dw/dT}\,\rho c_p$$

where dw/dT is the rate of change of saturation vapour density with temperature and ρc_p the thermal capacity of unit volume of air. Equating these quantities, we obtain

$$\frac{dr}{dt} = \frac{R}{4(L + \rho c_p/(dw/dT))}$$

With R = 8 mW cm^{-2}, L= 600 cal.cm^{-3}, ρ = 1/800 gm.cm^{-3}, C_p = 0.24 cal. $gm.^{-1}$ $^oC^{-1}$ and $dw/dT = 0.5 \times 10^{-6}$ gm.cm^{-3} $^oC^{-1}$. We find

$$dr/dt \simeq 0.4 \times 10^{-7} \text{ cm./sec.}$$

i.e. a rate of growth of 1 micron in about 4 minutes. The visibility in fog is roughly inversely proportional to the square of the drop radius, so that once the radius has reached the critical value of about 1.5μ a ten-fold fall in visibility (i.e. drop growth to about 5 μ) should occur in 14 minutes, in the postulated conditions. The radiative cooling mechanism thus seems quite capable of explaining the sudden increase in density of fogs so often observed.

The only obvious limit to the growth of fog drops by this method is set by the cutting off of radiation by the drops themselves. Presumably the fog should appear throughout the layer in which the relative humidity is high enough. As soon as the fog is dense enough to obscure the sky from the ground and the drops near it, the ground surface temperature will rise and the loss of heat by radiation from the drops near the ground will become very small. The final value of visibility reached near the ground will probably depend on a balance between the rate of growth of the individual drops and the rate of growth upwards of the fog as a whole. These rates depend on the size of the drop nuclei and on the temperature and humidity structure of the lower layers, respectively, so that the problem of predicting visibility is likely to be complicated.

APPENDIX VI

Variation of visibility with height

The observations plotted in fig. 11 suggest that fog is commonly thicker above the ground than at ground level. This appendix suggests a possible interpretation.

The upper surface of a well-established fog loses heat by radiation in much the same way as the ground surface on a clear night. This cooling will tend to make fog convectively unstable and thus set up convection currents conveying heat upwards. This is confirmed by the Cardington observations (figs. 3 and 4) which usually show a lapse rate close to the saturated adiabatic, in a deep fog. Within a fog, then, we must imagine that there are individual parcels of air ascending and descending, with condensation occurring in the ascending parcels and evaporation in the descending ones. It is reasonable to suppose that this condensation occurs on drops already in existence, so that the number of drops per unit volume should be much the same throughout the depth of the fog, but their size should be greater near the top, and the visibility should be less there than at ground level. If we make the further simplifying assumption that all the drops are of the same size at any given height, we can calculate the change in size with height and estimate the effect on fog density.

For a fixed number of drops per unit volume the visibility, V, is approximately inversely proportional to their cross-sectional area, i.e. to $m^{-2/3}$, where m is the mass of liquid water per unit volume of air, so that if V_o and m_o are the visibility and liquid water content at ground level and δm is the extra water condensed by adiabatic ascent to a height h, the visibility at h, V_h, is given by

$$V_h = V_o \left(\frac{m_o}{m_o + \delta m} \right)^{2/3}$$

$$= V_o \left(\frac{m_o}{m_o + \Gamma_w h \, dw/dT} \right)^{2/3}$$

since $\delta m = \Gamma_w h \, dw/dT$ where Γ_w is the saturated adiabatic lapse rate and dw/dT the rate of change of vapour density with temperature. Table VI shows values of V_h/V_o calculated from this formula for various values of h and m_o assuming $\Gamma_w = 0.004^{\circ}F/ft.$ and $dw/dT = 0.2$ gm./$m^3/^{\circ}F$ (the value appropriate to $40^{\circ}F$).

Table VI

V_h/V_o

H (ft.) \ m_o	0.1	0.2	0.5	1.0 g./m^3
0	1.00	1.00	1.00	1.00
100	0.68	0.80	0.91	0.95
200	0.53	0.68	0.83	0.93
500	0.35	0.47	0.68	0.80
1000	0.23	0.37	0.53	0.68

According to the table, visibility should decrease with increasing height, rapidly near the ground, more slowly higher up; the decrease is least marked when the water content is high, i.e. broadly speaking, when the fog is thick. This behaviour accords quite well, qualitatively and as to order of magnitude, with the observations of slant visibility plotted in fig. 11. It is impossible to make a quantitative comparison, because no measurements of liquid water content (m_o) were made.

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/TABLE II

TABLE II

Analysis of heat flow on selected occasions Units for columns (1) - (9) are milliwatts per square centimetre

Date and time	Height (ft.)	(1) Flux of radiation Up	(2) Flux of radiation Down	(3) Flux of radiation Net (Up)	(4) Radiative cooling of layer	(5) Observed cooling of layer	(6) Convective cooling of layer	(7) Convective flux (down)	(8) Latent heat of dew	(9) Flux of heat from earth	(10) Gradient of potl. temp. (°C /Km.)	(11) Eddy diffusivity (cm^2/sec.)	(12) Ri	(13) K_{RM} (cm^2/sec.)
27. 9.53	3000	35.7	26.2	9.5				0			5.5	0		
0100 hr.	1000	35.9	28.7	7.2	2.3	2.7	0.4	0.4			8.0	4,000		
	250	34.5	29.0	5.5	1.7	3.4	1.7	2.1			40.8	4,100		
	100	33.8	28.7	5.1	0.4	0.9	0.5	2.6			51.8	3,800		
	0	32.7	27.8	4.9	0.2	0.6	0.4	3.0	0.4	3.9	122+	1,900+	3.6+	370+
5.12.52	2250	30.4	21.6	8.8				0.9			1.1	66,000		
0001 hr.	1000	30.5	22.9	7.6	1.2	0.3	-0.9	0			3.6	0		
	500	30.3	23.2	7.1	0.5	0.6	0.1	0.1			17.9	450		
	100	29.5	22.6	6.9	0.2	0.9	0.7	0.8			67	930		
	30	29.2	22.5	6.7	0.2	0.2	0.0	0.8			69	920	0.24	13,000
	0	28.9	22.4	6.5	0.2	0.1	-0.1	0.7	0.4*	1.0				
28. 9.51	3000	35.0	25.0	10.0				0.5			3.5	11,000		
0001 hr.	2000	35.5	26.2	9.3	0.7	0.2	-0.5	0			2.6	0		
	100	34.4	27.4	7.0	2.3	4.4	2.1	2.1			51	3,300		
	0	33.6	26.9	6.7	0.3	0.6	0.3	2.4	0.7*	5.0	55+	3,000+	0.50+	4,500
31.10.53	3000	32.5	21.9	10.6				5.3			1.8	233,000		
0001 hr.	1000	33.3	24.9	8.4	2.2	-0.4	-2.6	2.7			6.2	35,000		
	0	31.0	24.6	6.4	2.0	1.4	-0.6	2.1	0.7	3.6	60+	2,800+	0.20+	14,000+

*Estimated from eye observations of dew +Values for 50 ft. level

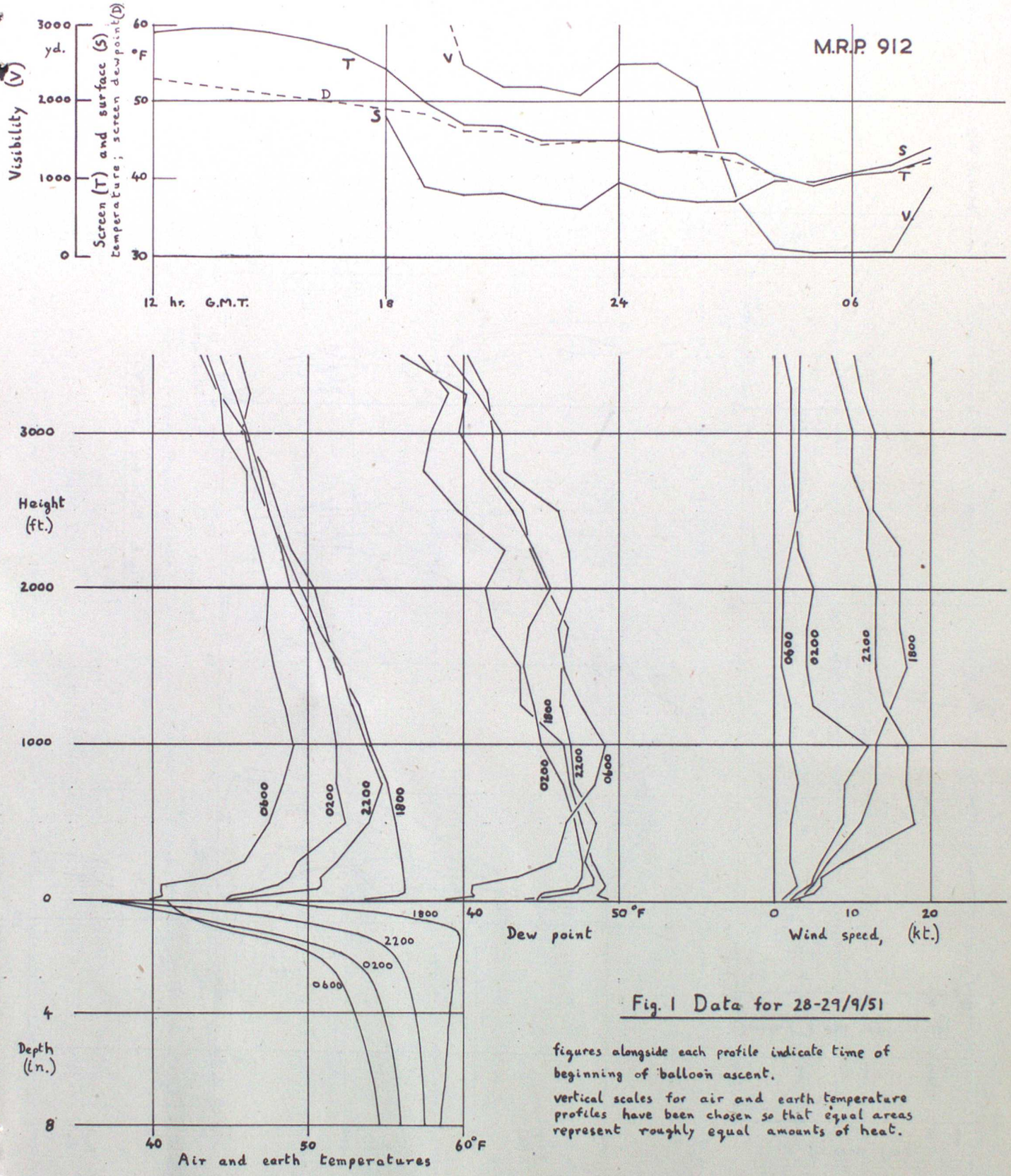
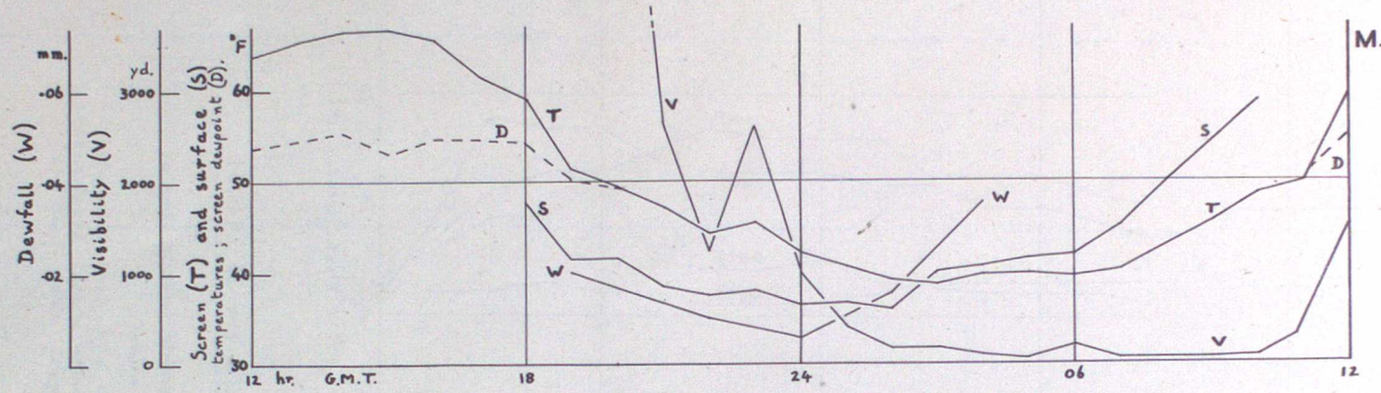


Fig. 1 Data for 28-29/9/51

figures alongside each profile indicate time of beginning of balloon ascent.

vertical scales for air and earth temperature profiles have been chosen so that equal areas represent roughly equal amounts of heat.



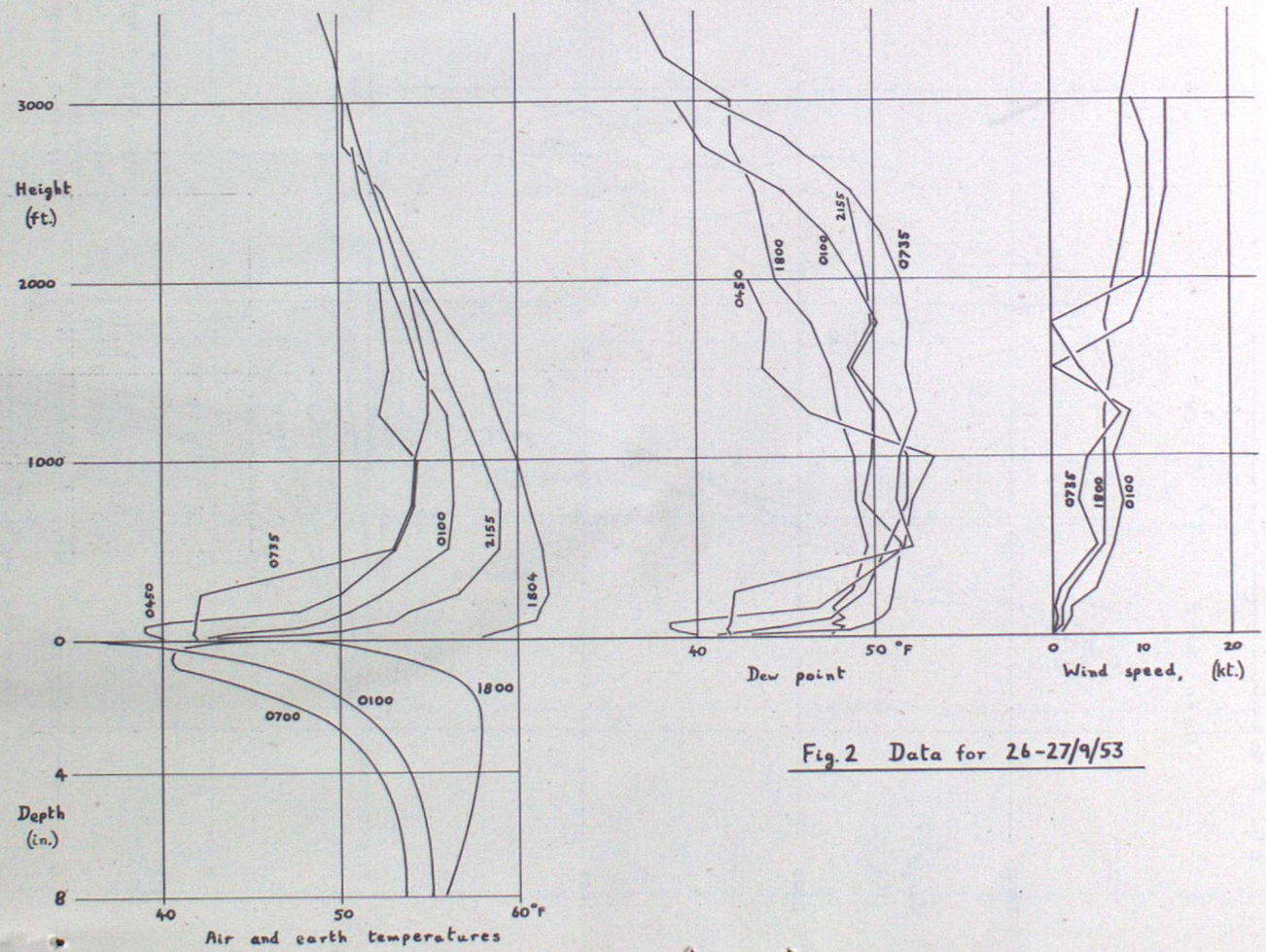
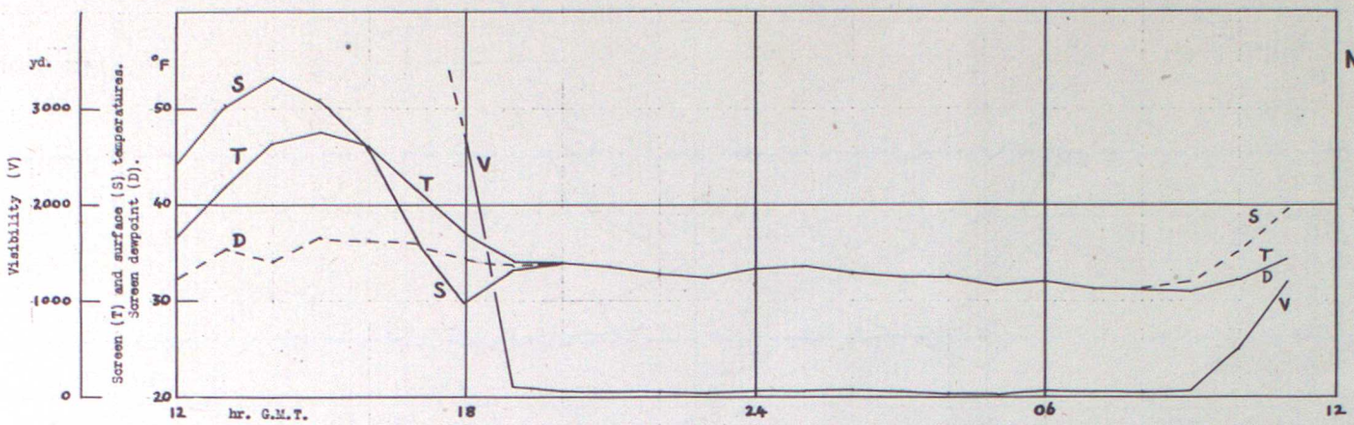


Fig. 2 Data for 26-27/9/53

M.R.P. 912



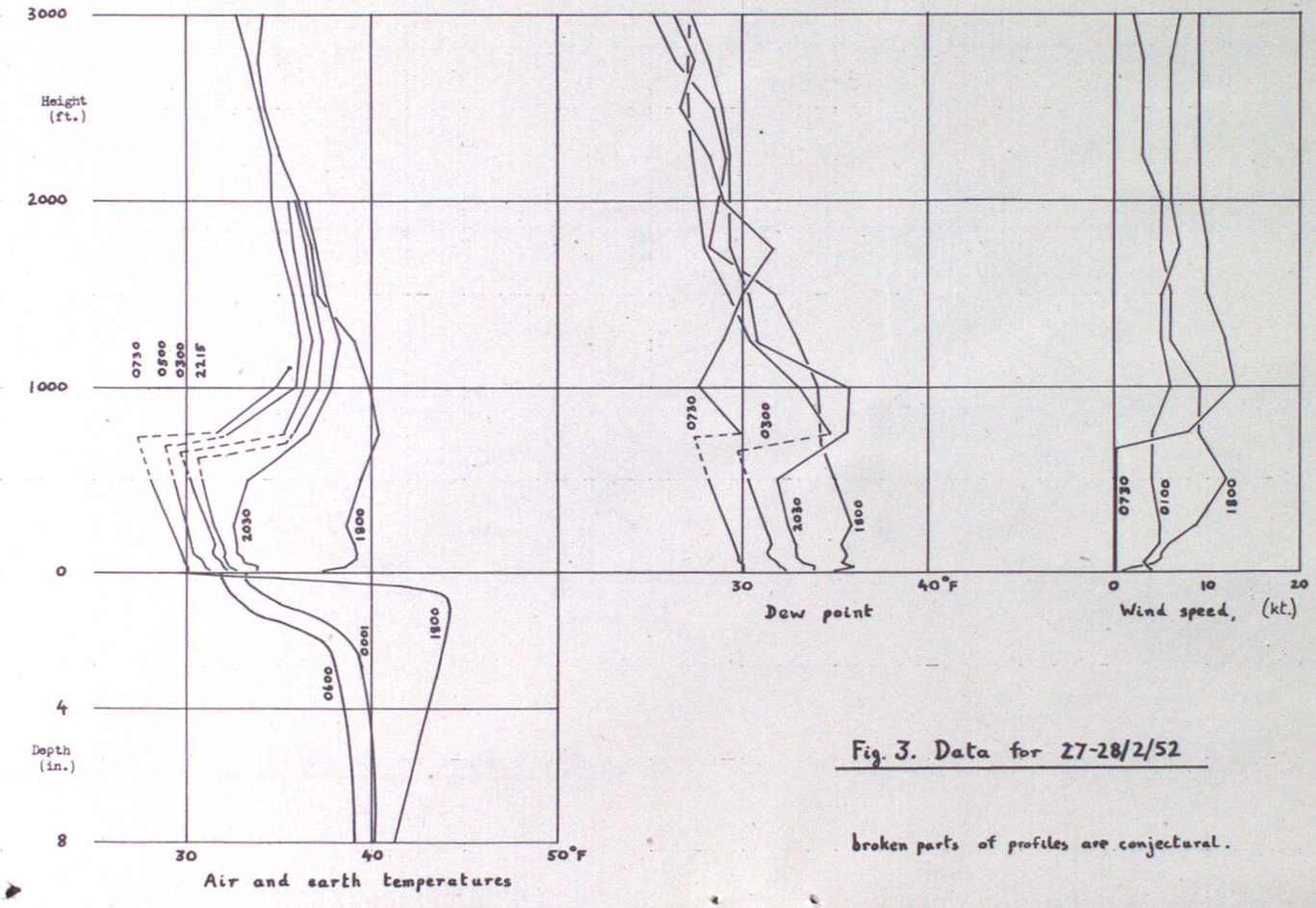


Fig. 3. Data for 27-28/2/52

broken parts of profiles are conjectural.

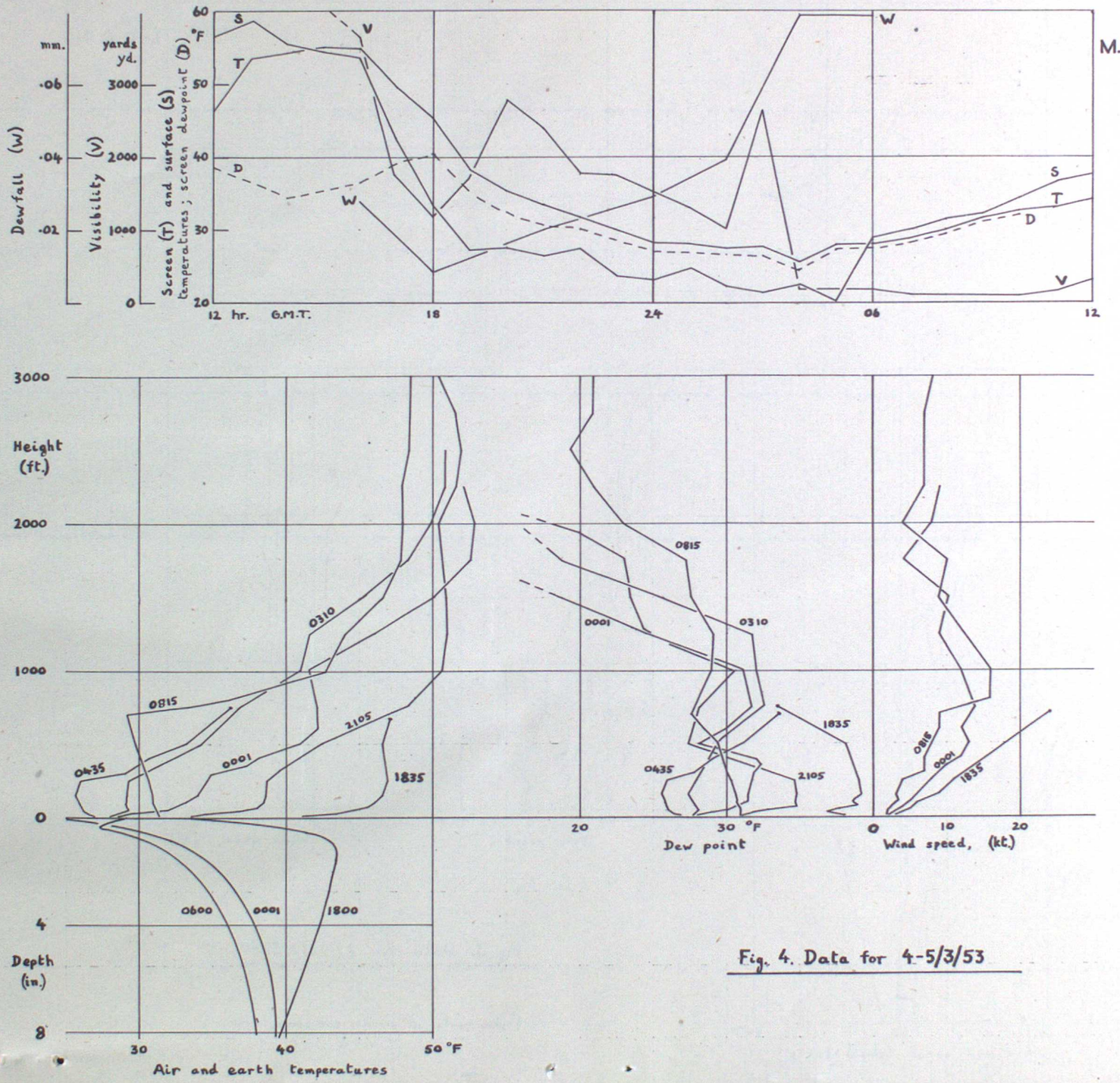


Fig. 4. Data for 4-5/3/53

M.R.P. 912

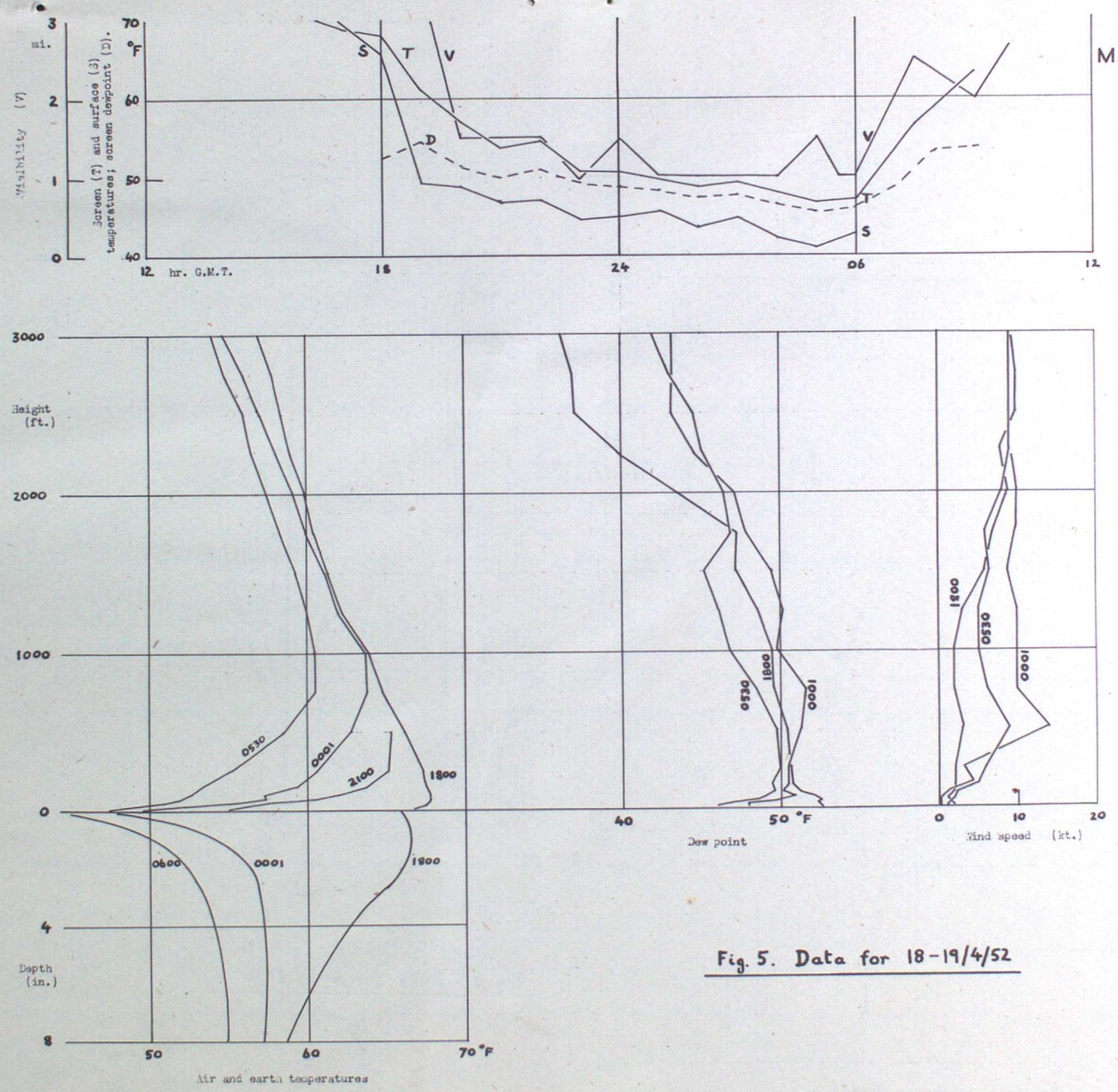
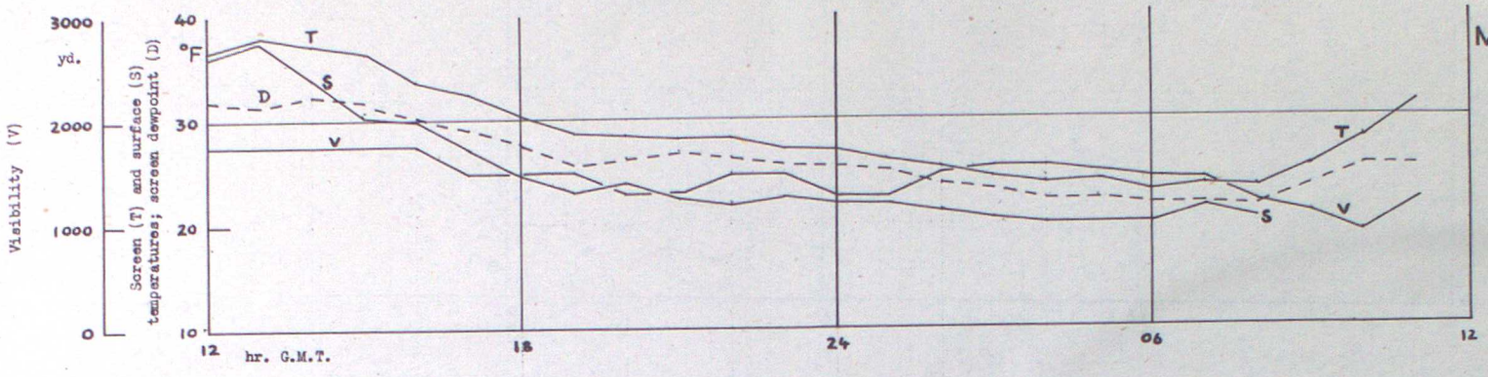


Fig. 5. Data for 18-19/4/52



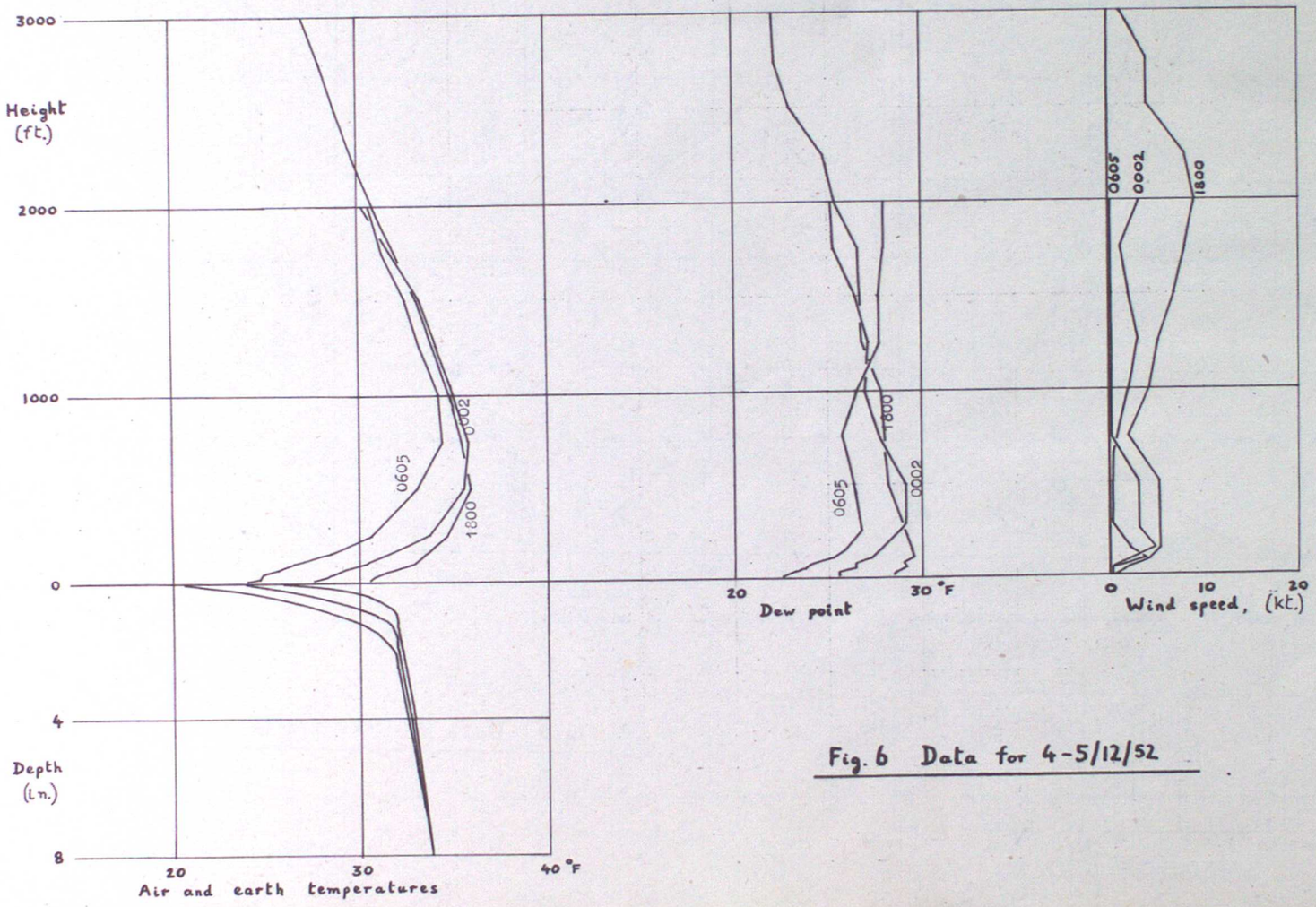


Fig. 6 Data for 4-5/12/52

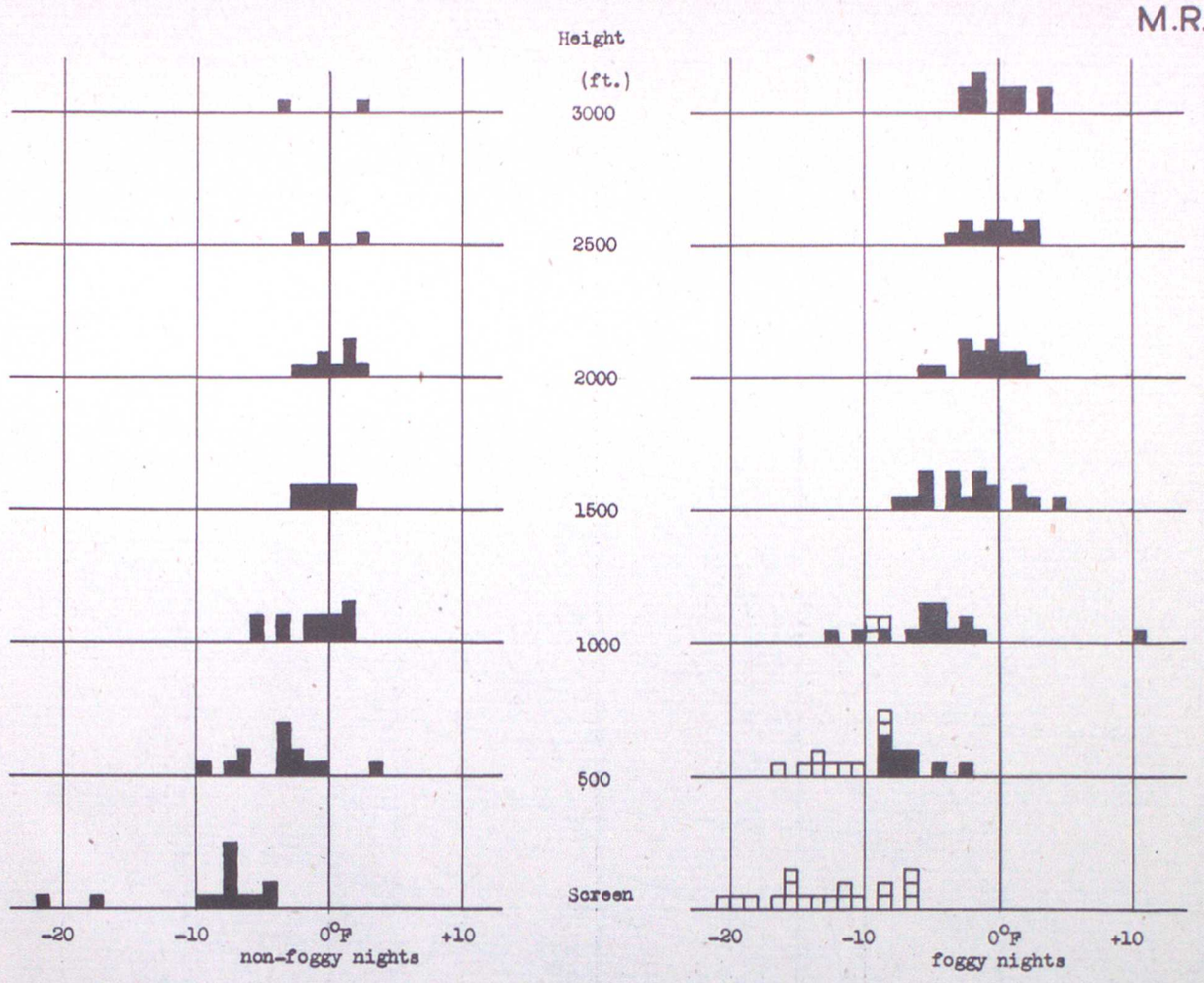


Fig. 7(a) Temperature changes 1800 - 0600 hr. Each square represents one occasion.
Open squares indicate observations within fog layer.

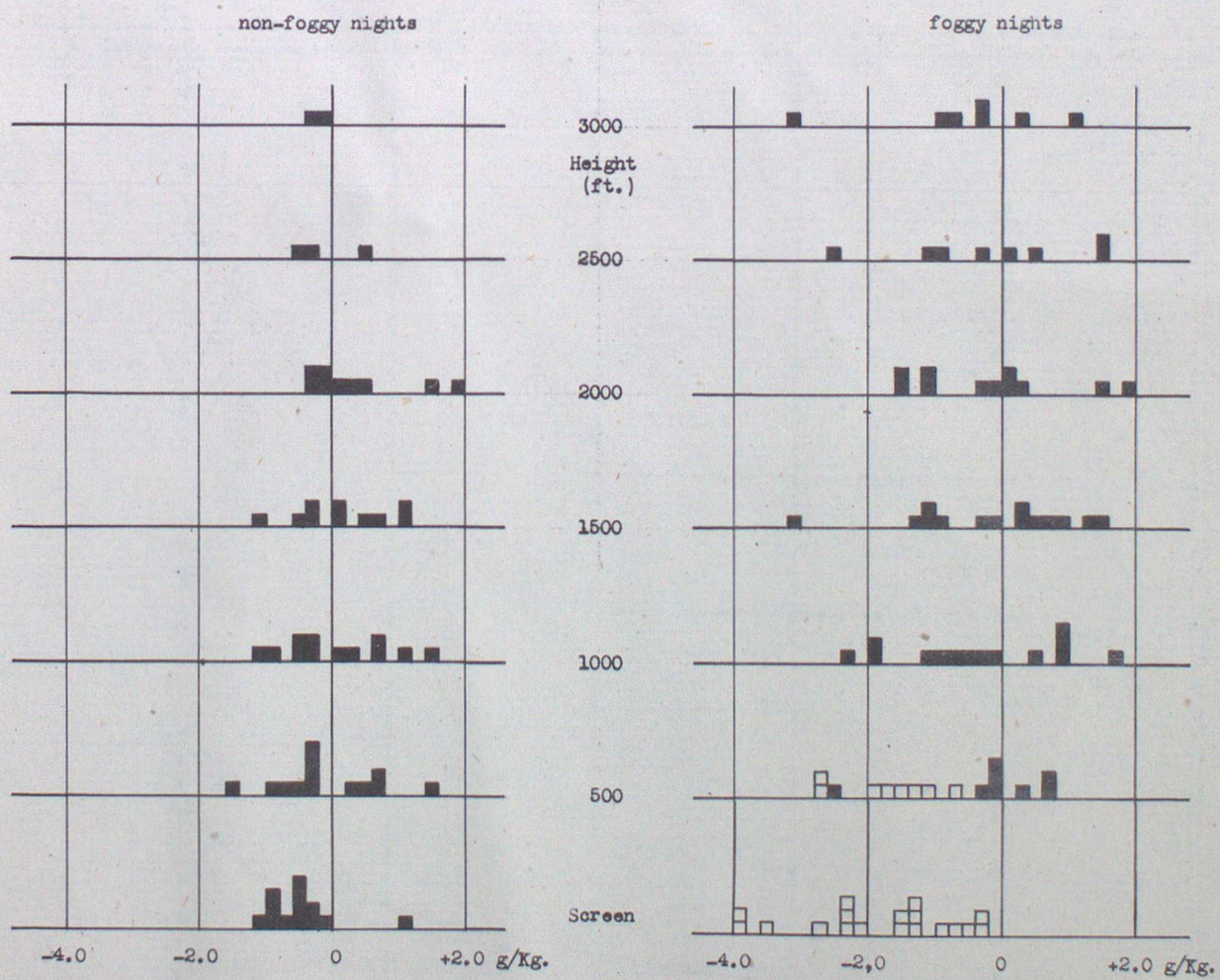


Fig. 7(b) Changes in specific humidity, 1800 - 0600 hr. Each square represents one occasion.
Open squares indicate observations in fog layer.

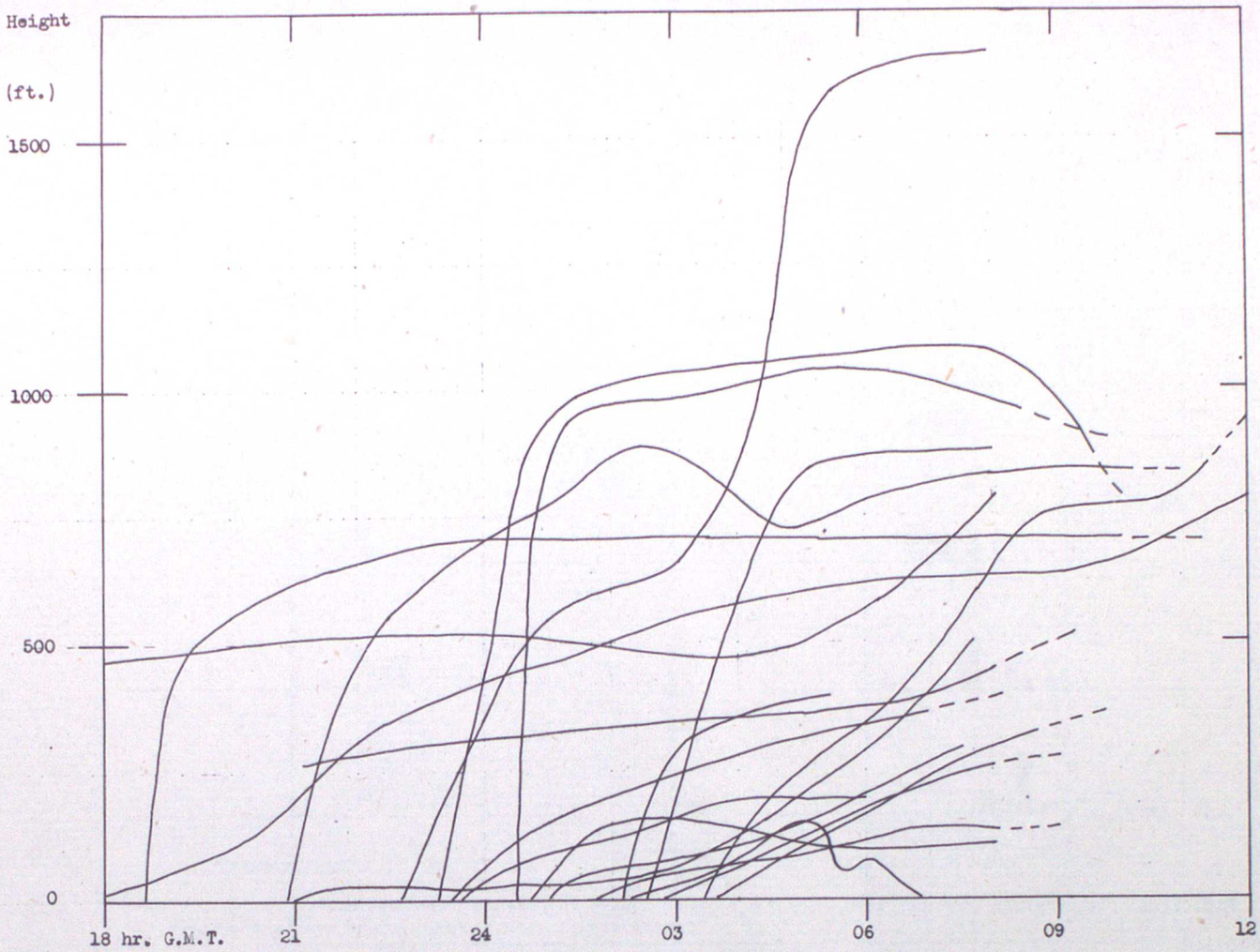


Fig. 8 Height of top of fog vs. time . Each line represents one night. Observations were mostly made at 2-3 hr. intervals and at standard heights, so curves are not very precise. Broken lines indicate dispersal of fog.

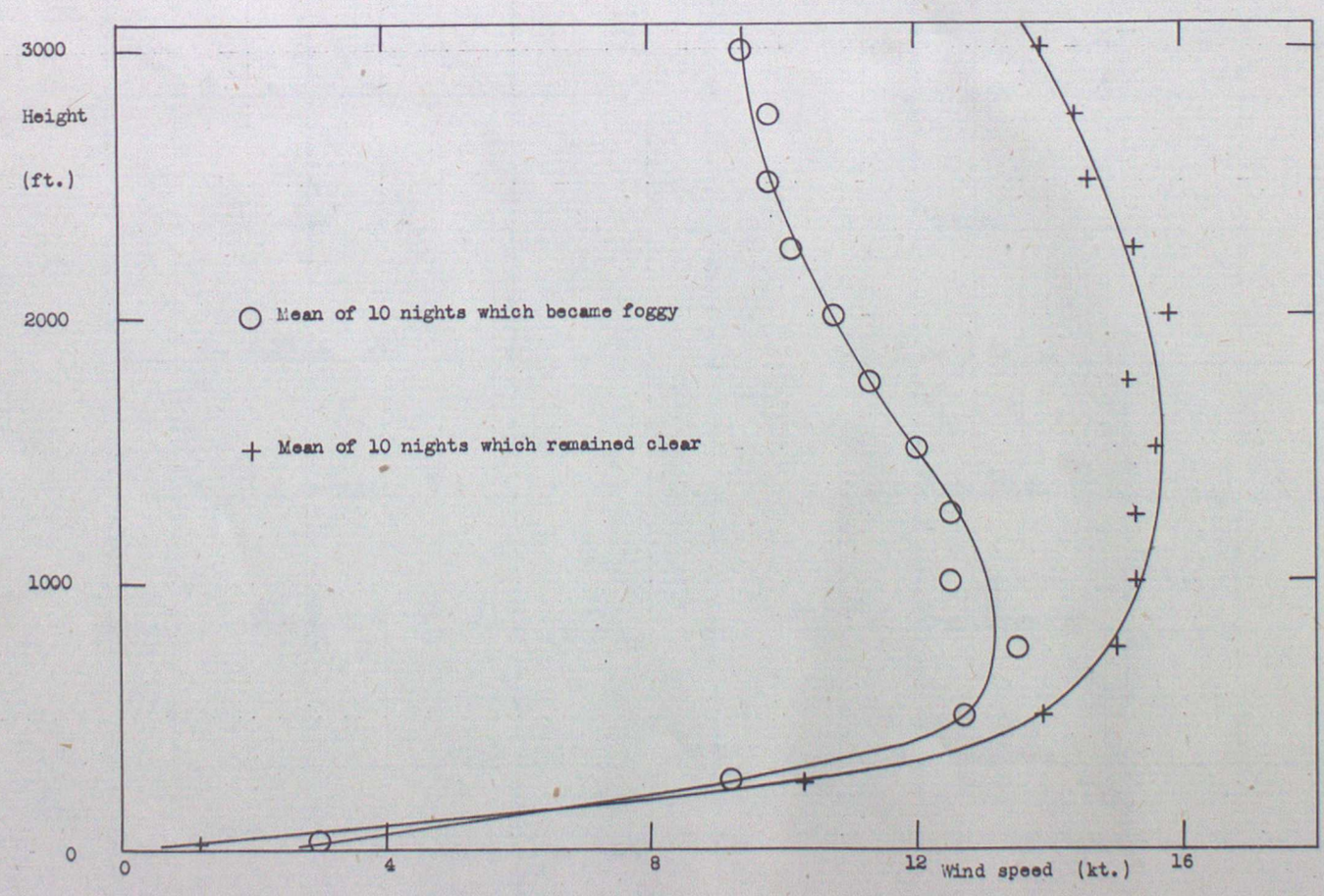
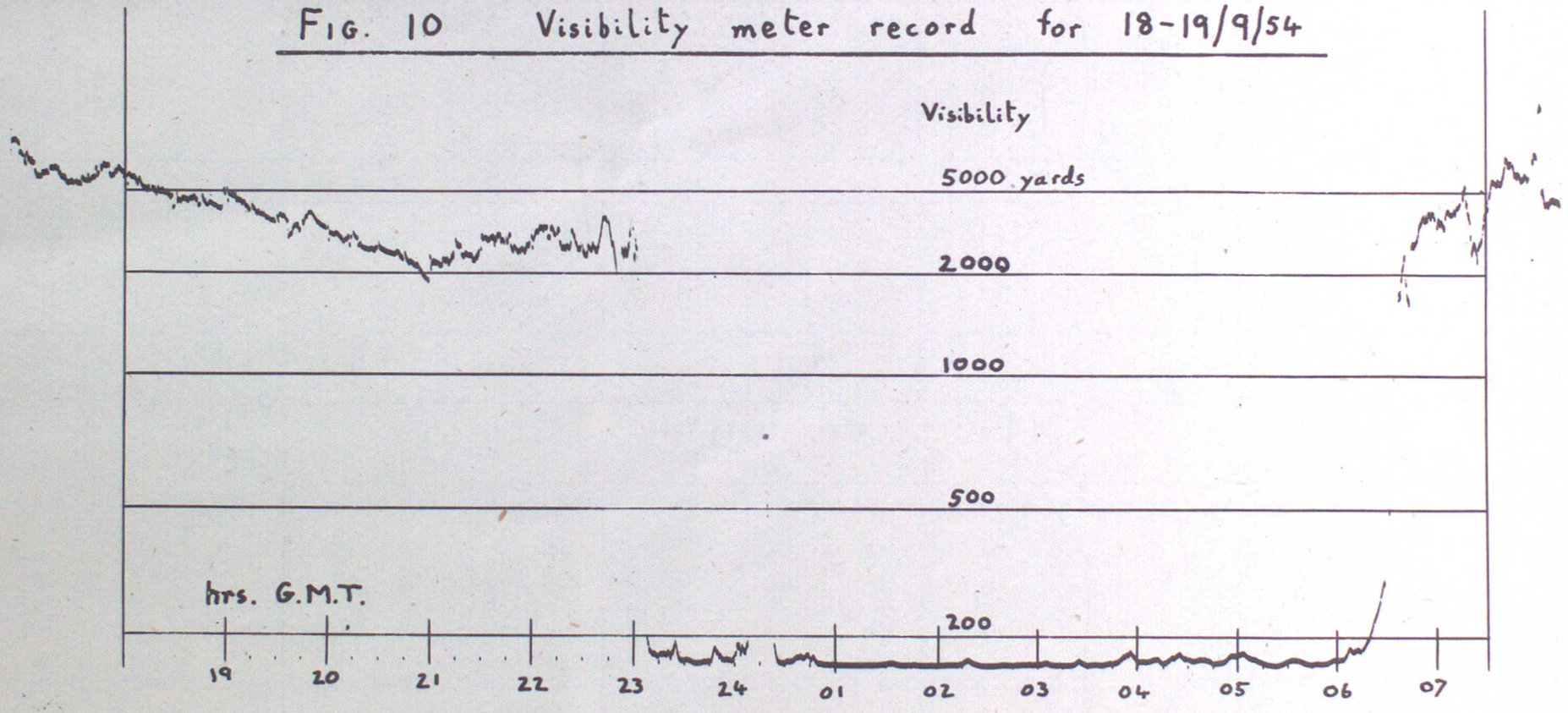


Fig. 9 Profiles of wind speed at Cardington at 1800 hr. on clear evenings

FIG. 10 Visibility meter record for 18-19/9/54



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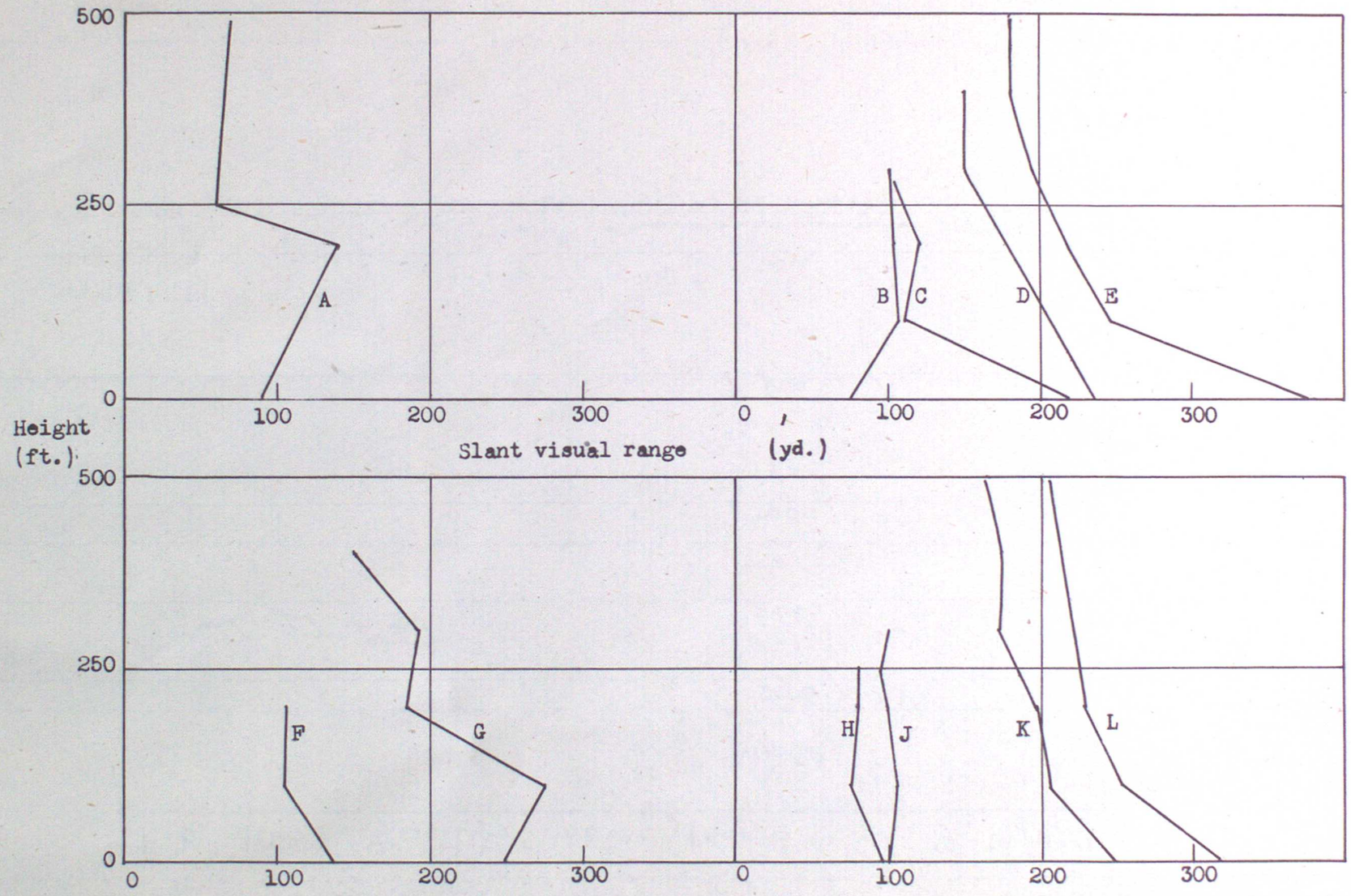


Fig. 11 Slant visual range of a 12 watt lamp vs. height for various occasions.

A 28/2/53 0130 B 3/3/53 0020 C 2/3/53 2140 D 1/3/53 2300 E 2/3/53 0500 F 3/3/53 0210
G 3/3/53 0500 H 4/3/53 0030 J 4/3/53 0245 K 21/3/53 0200 L 21/3/53 0400

Fig. 12

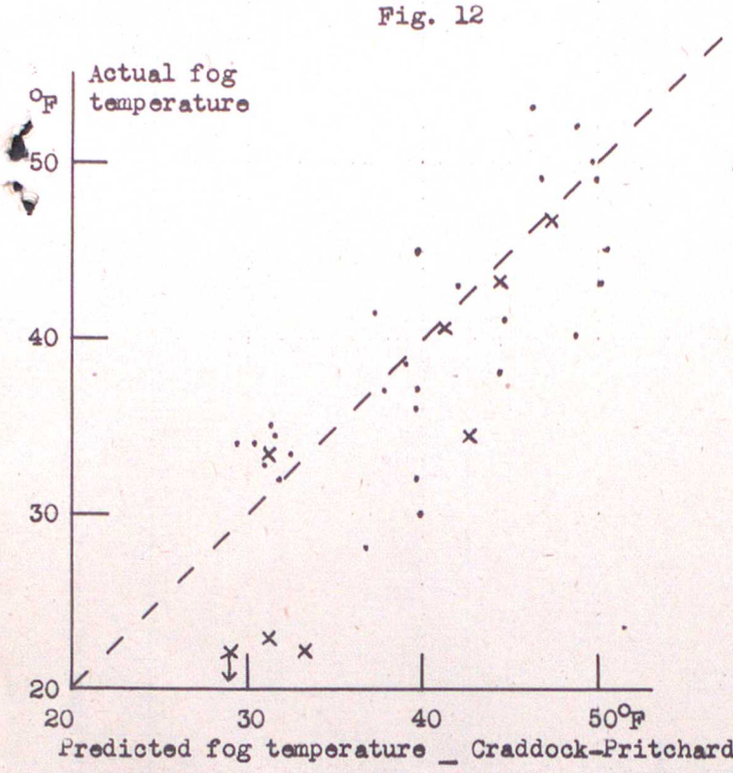
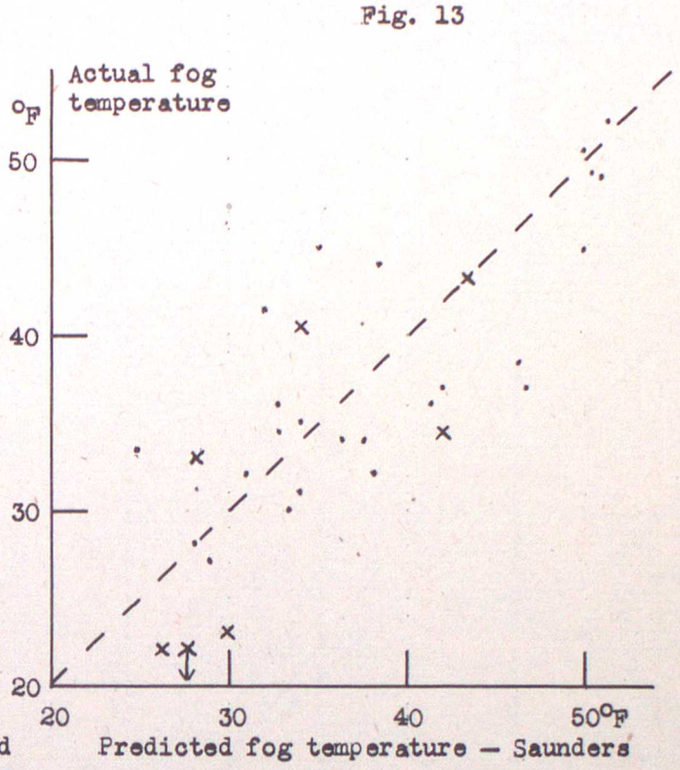


Fig. 13



Crosses represent cases where fog did not form; minimum temperature is plotted in place of "actual fog temperature".

Fig. 14

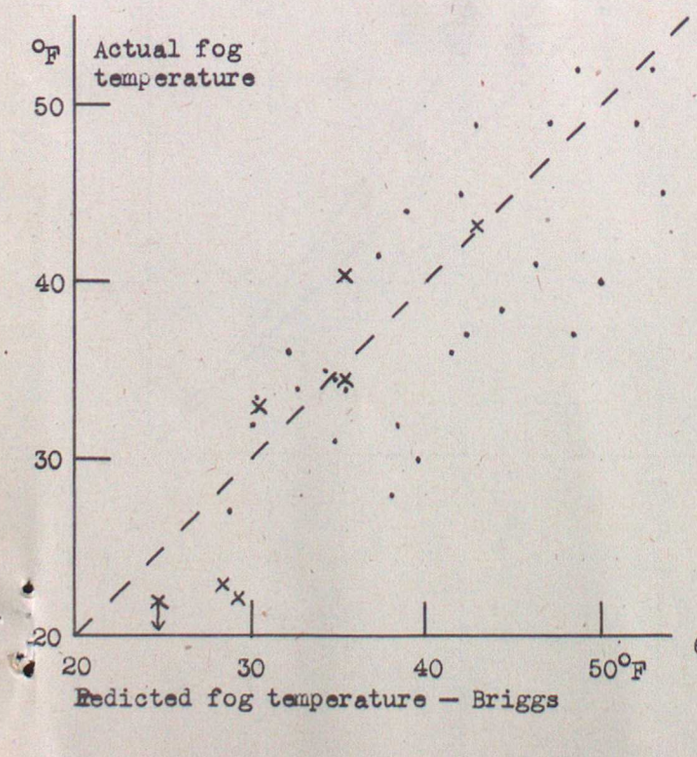
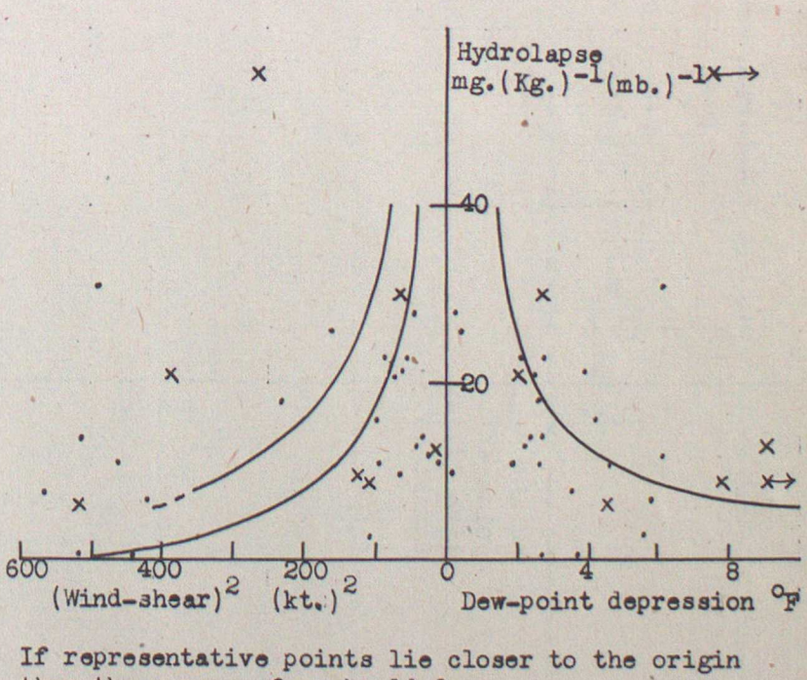


Fig. 15

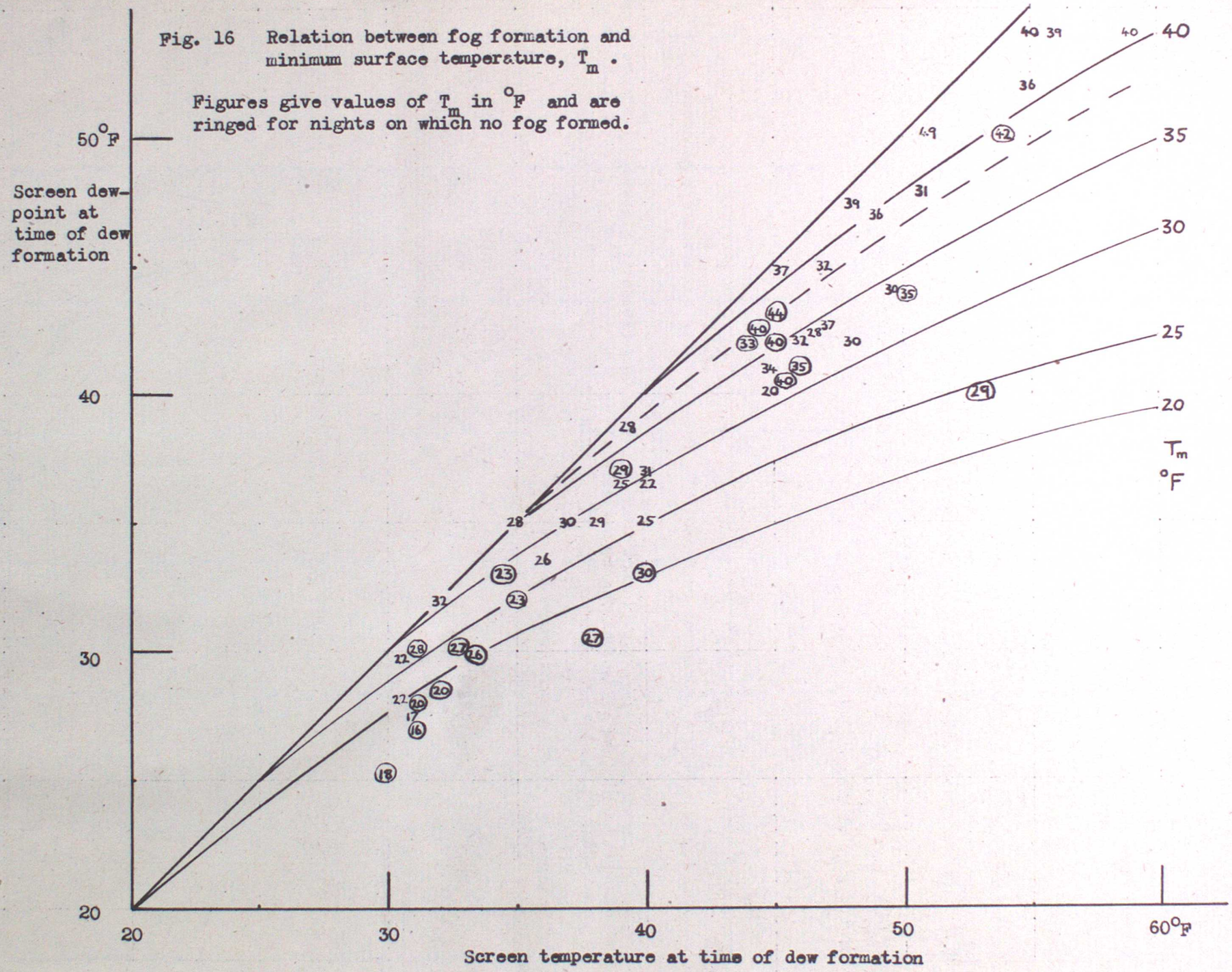
Swinbank's fog-prediction diagram



If representative points lie closer to the origin than the curves, fog should form.

Fig. 16 Relation between fog formation and minimum surface temperature, T_m .

Figures give values of T_m in °F and are ringed for nights on which no fog formed.



M.R.P. 912

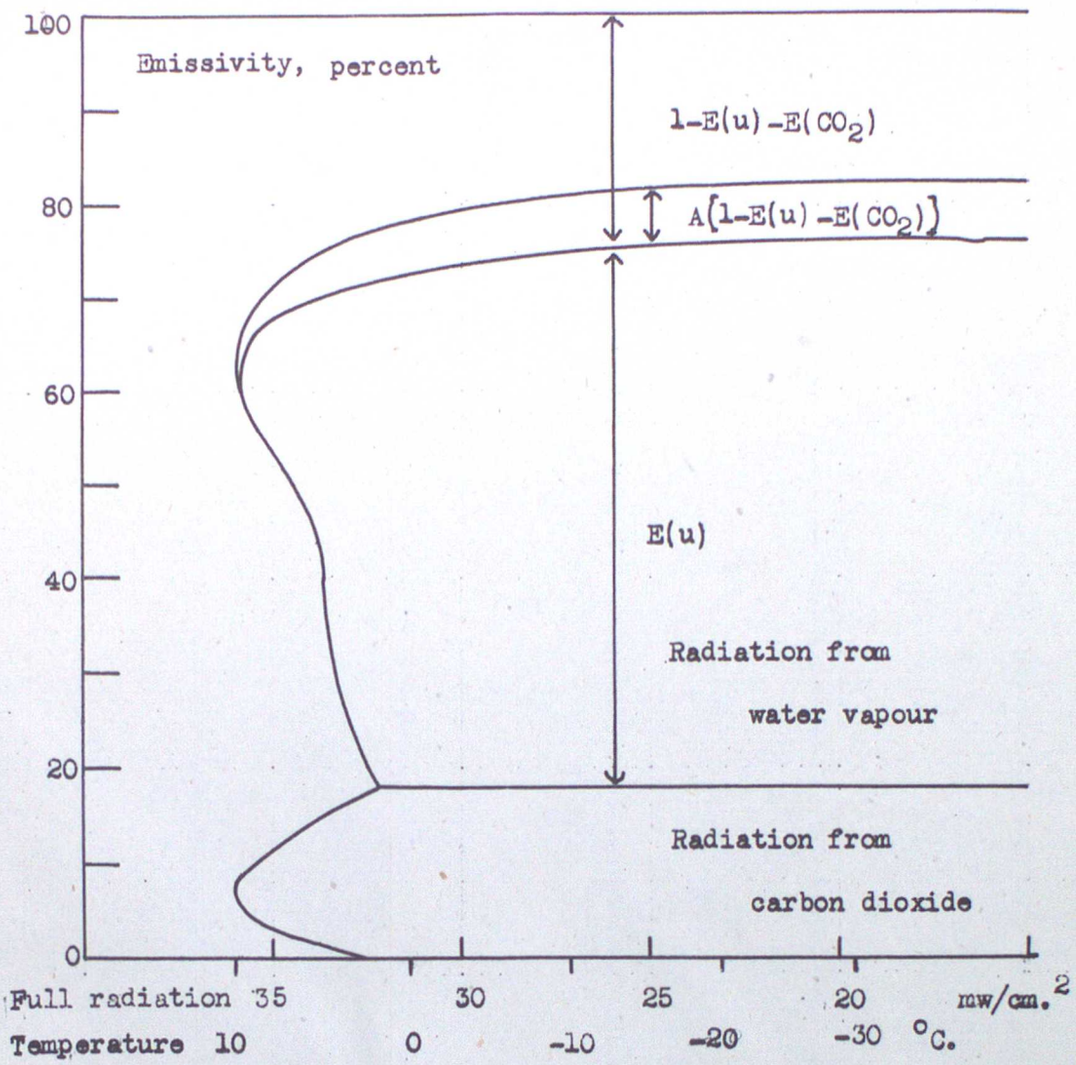


Fig. 17. "Kew" radiation chart for 0001hr. on 29/9/51, showing methods of estimating radiation from CO_2 and from particles.

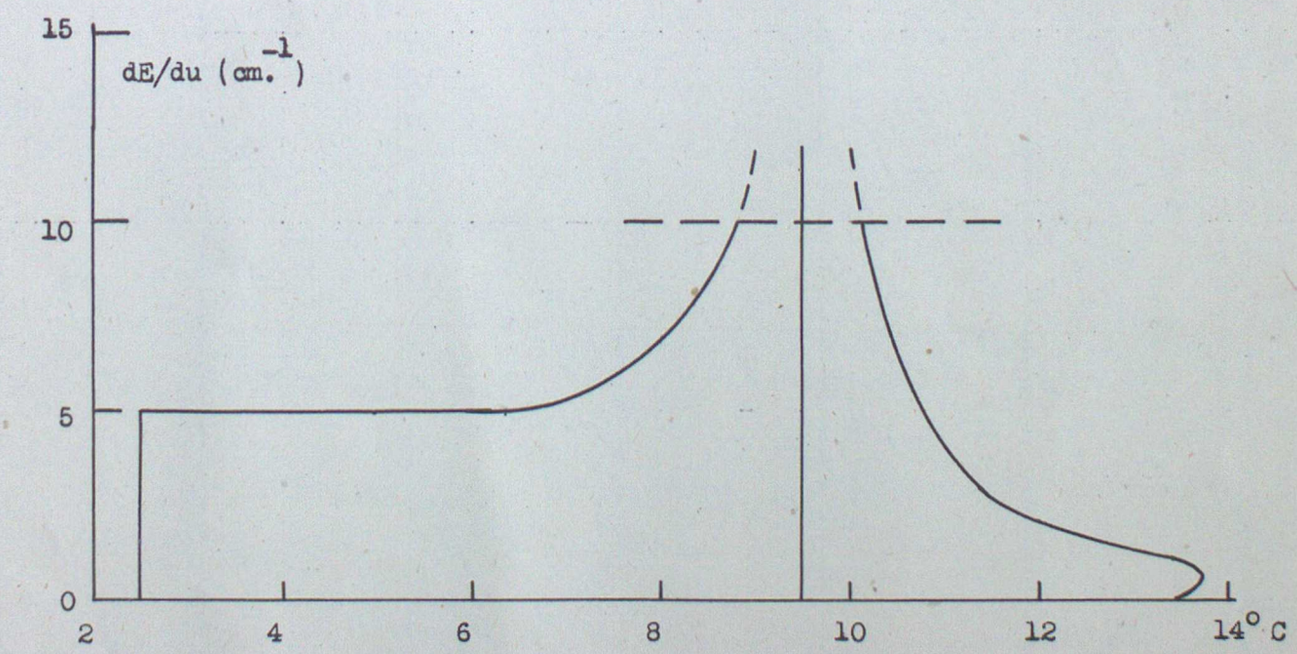


Fig. 18 Flux divergence chart for 100ft level at 0100 hr on 27/9/53. T is plotted as abscissa rather than σT^4 because σT^4 is roughly proportional to T over a limited range of temperature.